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PROCEEDINGS
WESTERN LIMITS OF DETACHMENT AND RELATED
STRUCTURES IN THE APPALACHIAN FORELAND

Southeastern Section, Geographic Society of America
Chattanooga, TN, April 1978

Russell L. Wheeler and Claude S. Dean, Editors

October 1980

UNITED STATES DEPARTMENT OF ENERGY
Morgantown Energy Technology Center
Morgantown, West Virginia

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INTRODUCTION

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On April 6, 1978, a symposium on "Western Limits of Detachment and Related Structures in the Appalachian Foreland" was held at the meeting of the Southeastern Section of the Geological Society of America, in Chattanooga, Tennessee. The purpose of the symposium was to furnish results of recent work to contractors of the Eastern Gas Shales Project (EGSP), and to explorationists generally. The work summarized dealt with fracture systems related to detached (thin-skinned) structures, at or near the western limit of detachment in the Appalachian Basin. Most but not all papers focused on the fractured Devonian shaly sequence that is the subject of EGSP's work.

The papers in this volume are published as received, without review, from camera-ready copy, and constitute the proceedings of that symposium. Some pertinent papers not presented at the symposium are also included. We have added footnotes to indicate address changes, citations of published papers that were in press or in review when copy was received, etc. To maximize exposure and dissemination of ideas, we have encouraged authors to publish elsewhere also. Where we know of such other publications, they are listed in footnotes.

After two general papers, the others follow in northeast-to-southwest order. We hope that the findings, data, and ideas reported in this volume can aid exploration for fractured gas reservoirs, both near the western limit of detachment, and as exploration expands eastward from present production into areas where the Devonian shaly sequence is deeper and more likely to be detached. These papers may also be useful in exploration elsewhere and in rocks of other than Devonian age, in the Appalachian and other overthrust belts.

The personnel, contractors, and coworkers of EGSP have published much additional information on fractured reservoirs, mostly in shaly rocks of Devonian age in the eastern United States. Short summaries of the problems attacked are given by Wheeler and others (1976) and Aguilera and van Poolen (1979). Details are in proceedings or preprints of annual EGSP meetings: Shumaker and Overbey (1976), Schott and others (1977) , Anonymous (1978), and Barlow (1979). A list of open-file material is available from Dorothy Simon, Librarian, Eastern Gas Shales Project, Morgantown Energy Technology Center, U. S. Department of Energy, P. 0. Box 880, Morgantown, WV 26505. Other material has been and is being published continually in journals and by EGSP.

INTRODUCTION

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POSSIBLE INTERACTION BETWEEN THIN-SKINNED AND BASEMENT TECTONICS IN THE APPALACHIAN BASIN AND ITS BEARING ON EXPLORATION FOR FRACTURED RESERVOIRS IN THE DEVONIAN SHALE

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ABSTRACT

The Devonian shale of the Appalachian Basin is an enormous, virtually untapped resource of natural gas. Production rates are low; even better wells are marginally economic. Most of the gas produced arrives at the wellbore via natural fractures; the direct contribution from the shale matrix, where most of the gas is held, is minimal. Hence, exploration for highly-fractured areas within the shale is a most important aspect of the Department of Energy's Eastern Gas Shales Program. Fracture permeability is primarily the net result of jointing that occurred following consolidation. Jointing ideally occurs whenever fluid pore pressure exceeds the minimum compressive stress (σ_3) by the tensile strength (K) of the rock, provided σ_1 - a3 $\leq$ 4K. Consequently, the formation of joints, which show evidence of being tension fractures, is not depth restricted (Secor, 1965). Carbonate mineralization of many natural fractures seen in Devonian shale core testifies to the role of fluids. Thus, joints, originating as natural hydraulic fractures, are oriented perpendicular to the prevailing σ_3 and provide a cumulative record of stress history. Were they exposed, *it* should be possible to unravel the tectonic history of the shale by establishing a relative chronology among sets. In practice, the explorationist must reverse the process. Using the tectonic histories of marginal areas and the underlying basement as boundary conditions, he is forced to synthesize a tectonic history for the Devonian shale and utilize it to predict ensuing joint patterns, from which he hopes to locate concentrated fracturing. Finite element stress analysis is a powerful technique for calculating stress trajectories and intensities, but ignorance of tectonic influences severely limits its application. Of particular significance are possible interactions between thin-skinned and basement tectonics.

INTRODUCTION

The purpose of this paper is threefold:

1. To introduce the reader to the U.S. Department of Energy's (DOE) Eastern Gas Shales Project (EGSP), which is the tacit incentive for this symposium and under the auspices of which these proceedings are being published.
2. To offer some fundamental insights on jointing in the Devonian shale from the point of view of *rock* mechanics.
3. To discuss how a growing understanding of the thin-skinned and basement tectonics of the Appalachian Basin should influence exploration for natural gas from the Devonian shale.

The common thread that unifies these seemingly disparate topics is natural fracture systems in the Devonian shale, the presence of which the authors judge to be critical to commercial natural gas production.

DEVONIAN SHALE

Dwindling reserves of natural gas are an important aspect of the nation's energy crisis, but one, commonly overlooked with the current focus on scarce oil supplies. Yet, natural gas currently supplies about 30 percent of our national energy needs. Yearly consumption is about 20 tcf (trillion cubic feet). At projected rates of gas discovery, demand could outstrip domestic supply by 10 tcf per year or more by 1990. Consequently, unconventional resources of natural gas are receiving closer scrutiny than ever before. One such is the Devonian shale of the Appalachian, Illinois, and Michigan basins.

"Devonian shale" is the term drillers employ to refer to all fine clastic strata intervening between the lower Middle Devonian Onondaga Limestone and the Lower Mississippian Berea Sandstone, or their equivalents. Where it overlies the Berea, the dark Sunbury Shale is also included. Thus, the term comprises the bulk of the Devonian section in all three basins and includes some lower Mississippian strata as well. Figure 1 shows the wide distribution of Devonian shale across the eastern half of the United States. The thickness of the comprised section ranges within the Appalachian Basin from less than 1000 feet in the west to more than 7000 feet in the east; it is on the order of a few hundred feet in the other two basins. Because of their capacity to generate natural gas under favorable geological conditions, the dark, organic-rich shales are of particular economic interest; though, over much of the Appalachian Basin gray shale and coarser clastics dominate the section. The proportion of dark, organic shales varies from over 80 percent in the western basins to about 10 percent in the eastern portion of the Appalachian Basin. Where it crops out the Devonian shale has historically been subdivided into named stratigraphic units. Some of the more familiar names are Hamilton, Marcellus, Chattanooga, Ohio, New Albany, Antrim, etc.

RESOURCE POTENTIAL

Despite its novelty as an important unconventional resource of natural gas, commercial Devonian shale gas production is as old as the industry itself. Fredonia, NY, 1821, the first gas well drilled in the United States produced from Devonian shale at a depth of 27 feet. Subsequently, natural gas for local consumption has been profitably extracted from hundreds of Devonian shale wells drilled to depths of a few hundred feet along the southern shore of Lake Erie. In the more than 150 years that have elapsed since the drilling of the first Devonian shale gas well, commercial gas production from Devonian shale has been established from scattered areas in all three basins. Big Sandy, the largest Devonian shale gas field (also the largest gas field in the eastern United States), covers most of eastern Kentucky and includes parts of West Virginia, Virginia, and Ohio.

Yet, natural gas from Devonian shale constitutes but the minutest fraction of the total consumed in the United States. Commercial gas producers have never considered the shale to be a prime drilling target to be compared with the traditional sandstone and carbonate reservoirs, but only marginally economic at best. The reason becomes evident upon comparing production decline curves (figure 2). The rate of production from wells completed in conventional reservoirs is initially much higher, often by an order of magnitude, than from those completed in the Devonian shale. The rate of production decline, however, tends to be much less for Devonian shale wells than for sandstone or carbonate wells. Although cumulative production from Devonian shale wells may ultimately exceed that from conventional wells, the relatively long payout time and low annual return on investment makes the former economically unattractive relative to the latter. Regulated gas prices coupled with high inflation aggravate the situation.

Nevertheless, the Devonian shale is an enormous resource of natural gas, as yet virtually untapped. Up to 15 cubic feet of gas may be contained in a ton of black shale. Estimates of total gas-in-place vary from 300 to 900 tcf, and these may be conservative by a factor of two or three. More important and more difficult to estimate is the amount of gas economically recoverable from the Devonian shale. That depends on a number of largely imponderable factors, such as future price and unforeseen technological advances. Estimates have ranged widely from a very pessimistic 10 tcf to a very optimistic 500 tcf. A couple of hundred trillion cubic feet (tcf) would appear to be a reasonable compromise, but that is equivalent to a decade's gas supply at current consumption rates. Total historic production is estimated to have been about 3 tcf.

Expansion of Devonian shale gas reserves within the confines of the total resource is dependent on three factors: economics, technology, and geology (figure 3). Simply stated, economics equals price. The higher the wellhead price of gas, the greater will be the incentive to drill the

Devonian shale, even in the face of low production rates. Extraction technology is critical to exploiting the shale, as most Devonian shale wells produce negligible quantities of gas prior to stimulation. Geology or, more precisely, the exploration for environments favorable to the production of gas from the Devonian shale is the object of this paper and, indeed, this symposium.

The U.S. Department of Energy's (DOE) Eastern Gas Shales Project (EGSP) was initiated in 1976 and now continues as an integral part of its nationwide Unconventional Gas Recovery Program. The EGSP aims to promote further commercial development of natural gas supplies from the Devonian shale. It has two principal emphases that essentially correspond to two of the above-mentioned factors on which shale gas reserves depend: Resource Characterization and Evaluation (Geology) and Extraction Technology Development, Testing, and Verification (Technology). Responsibility for the third factor (Economics) falls outside the scope of the project; that belongs to the regulatory arm of DOE. Exploration Research and Development is an important ongoing activity under Resource Characterization and Evaluation. It represents the practical application of resource quality information to the selection of optimum areas, and even sites, for extraction technology demonstration projects. The goal of this symposium is to illumine the tectonics of the Appalachian foreland in such a way as to reveal promising exploration rationales for the Devonian shale.

PRODUCTION FROM FRACTURED RESERVOIRS

In driller's parlance the Devonian shale is an extremely "tight" formation, meaning that natural gas does not easily flow to the wellbore as it does in some of the more permeable sandstone reservoirs. In fact, the matrix permeability of the shale is so low that it is measurable only in fractions of a microdarcy (Smith, 1978). R. D. Smith (personal communication) estimated that it takes perhaps 30 years or more for a molecule of gas to move through one centimeter of shale under the impetus of the pressure gradient existing in a typical Devonian shale well. Matrix permeability alone is thus not anywhere near sufficient to account for existing Devonian shale production. No known means of artificial stimulation is capable of inducing gas production from a formation with such a low matrix permeability, if that is the only avenue available for gas transport. Matrix permeability that low is presumptive evidence that where the shale produces natural gas the total formation permeability must be very greatly enhanced by natural fractures. Fractures provide additional surface area through which to drain large volumes of matrix, albeit slowly, and act as conduits leading eventually to the wellbore.

Other evidence and lines of reasoning also yield the conclusion that Devonian shale production is from fractured reservoirs. Figure 4 compares average production decline curves for Devonian shale wells differentiated according to open flow (i.e., initial open flow after stimulation). There is a high degree of variability in the performance of Devonian shale wells, but, a consistency in the pattern of decline. Wells with high initial production decline rapidly during the first several years of production, but then they progressively level off, eventually reaching a near steady-state with an imperceptible rate of decline. Wells with low initial production decline very little, but reach the near steady-state early in life. The higher the initial production, the higher will be the level of the near steady-state production. Devonian shale wells commonly display hyperbolic production decline, as opposed to the more conventional exponential decline. All this behavior is thoroughly consistent with P. J. Brown's (1976) model in which the total volume of gas in the Devonian shale (V_T) is held in three distinct ways:

1. V_1 = Free gas retained in fractures.
2. V_2 = Adsorbed gas on the walls of fractures.
3. V_3 = Absorbed gas within the shale matrix.

V_1 gas accounts for the initial "flush" production of Devonian shale wells, but it is soon replaced by slower devolving V_2 gas. Production of V_3 gas is limited by diffusion through the virtually impermeable shale matrix and it can only be produced on reaching the nearest fracture connected to the wellbore. V_3 gas constitutes by far the greatest proportion of V_T, as suggested by the near steady-state portion of the production decline curves.

To characterize the Devonian shale and evaluate its natural gas resource potential, the EGSP has undertaken an ambitious coring program in the Appalachian, Illinois, and Michigan basins. Fifty cores have been recovered to date. A few are highly fractured; most are only moderately so; some are virtually unfractured. Though much data remain to be compiled and analyzed, the cores do provide some unambiguous, direct evidence of the dependency of gas production on natural fractures in the shale.

In 1975 the EGSP and the Consolidated Gas Supply Corporation (CGSC) (see Martin and Nuckols, 1976) cooperatively cored two wells in the Cottageville (Mt. Alto) Devonian shale gas field in Jackson and Mason Counties, West Virginia (figure 5). The wells are located on opposite sides of the field: CGSC 11940 L. A. Baler to the south and CGSC 12041 W. L. Pinnell to the north. CGSC 11940 had a natural open flow of 1,100 MCFD (thousand cubic feet per day) and was completed as a natural producer, that is, without any form of artificial stimulation. CGSC 12041 had no discernible natural open flow and was completed as a marginal producer following a "large-scale foam fracture" treatment (Frohne, 1977), after which the absolute open flow potential was only 173 MCFD.

Figure 6 shows the section of the gamma ray log from CGSC 11940 that includes the cored intervals. The highly radioactive, organic-rich "Zone II" shale as designated by Martin and Nuckols (1976) (now known to correlate with the Lower Huron Shale of Ohio) is the primary producing interval in the Cottageville Field. Fracture orientation data, derived from the oriented core and displayed in rose diagram form, are correlated to the gamma ray log in figure 6. The relationship between fracture orientation, stratigraphic interval, and gas production is a striking one. The dominant fracture orientation in the core is N45E. Between 3700 feet and 3790 feet, however, the EGSP core logging team (Byrer, Trumbo, and Rhoades, 1976) recorded both a substantially greater number of fractures and a far greater diversity of orientations than in the rest of the core. It was this interval within "Zone II" that was responsible for the 1.1 MMCFD (million cubic feet per day) natural open flow from CGSC 11940.

At the time the CGSC 11940 core was processed, the practice was to log fractures indiscriminately on a per foot basis. On close examination of fractures in shale cores recovered subsequently in the program, the senior author and others (Kulander, Dean, and Barton, 1977) realized that many of the fractures were coring induced. There are several recognizable types of these coring induced fractures. One of the more common is a vertical fracture with a relatively smooth planar surface that tends to track down the center line of the core, bisecting it into two equal parts. This is the "centerline" portion of the "petal-centerline" coring induced fracture of EGSP usage (Kulander, Barton, and Dean, 1979, p. 134). This type of fracture proceeds into the core from the margin at an angle of about 45° from the vertical and curves downward ("petal" portion) to conform to the centerline, whence it continues downward by discrete extensions. Though the mechanics of petal-centerline fractures are complicated (GangaRao, Advani, Chang, Lee, and Dean, 1979), their orientation reflects the anisotropy of the modern in situ stress field and thus tends to be faithfully consistent within a given core. The authors are certain that most, if not all, of the N45E striking fractures are coring induced centerline fractures. The CGSC 11940 core is no longer intact, so this assertion cannot be proved directly. The indirect evidence, however, is overwhelming:

1. Uniform strike over an appreciable interval is characteristic of petal-centerline fractures.
2. Photographs of the core reveal unmistakable petal-centerline fractures. Figure 7 gives an edge view of one; the face of another is revealed in the left hand portion of figure 8.
3. A fracture strip log of the core shows diagrammatic representations of what can only be petal-centerline fractures throughout most of the cored interval.

The diversely oriented fractures in the 3700 to 3790 interval (figures 9 and 10), however, are certainly of geologic origin. Only within that interval did Byrer, et al. (1976) record mineralized (dolomite) fractures, an observation subsequently confirmed and expanded by Patchen and Larese (1976, p. 7) and by Larese and Heald (1977 , pp. 19-22) (see figures 11 and 12).

In contrast to the CGSC 11940 core, the CGSC 12041 core was virtually devoid of fractures. Those few that were observed all strike N65E and are probably coring induced centerline fractures. Thus, the two EGSP core wells in the Cottageville Field, less than 5 miles apart, serve as dramatic contrasting examples of the dependency of Devonian shale gas production on natural fractures, and especially intersecting, interconnected natural fracture systems.

ORIGIN OF NATURAL FRACTURES

What is the true nature of these all important natural fractures? Are they a highly localized phenomenon, in which case Devonian shale production will be forever confined to the few areas where fractures occur? Or, are they ubiquitous, variations in fracture density accounting for the distinction between traditional producing and nonproducing areas?

Outcropping dark, organic-rich Devonian shale is conspicuously jointed, typically displaying one or two systematic sets and a corresponding number of associated nonsystematic sets. Systematic joints of a given set are characteristically highly regular, i.e., evenly spaced, planar, and smooth. Visual comparison suggests that black shale units are more highly jointed, certainly more regularly and obviously jointed, than superjacent and subjacent gray shale units. The great majority of joints contained in a given black shale unit terminate at the upper and lower contacts with gray shale. (The senior author observed the foregoing on a field trip through central and western New York; Patchen and Dugolinsky, 1979.) Do these joints exist at depth? Are they the source of gas productive fracture permeability in the Devonian shale of the subsurface?

A common misconception, espoused by several respected textbooks, holds jointing to be a near-surface phenomenon, primarily the result of erosional unloading of buried rock masses. The reasoning generally goes as follows:

1. Joints are tension fractures, as indicated by several lines of evidence.
   a. Imperceptible offset of pre-existing geologic features.
   b. Tendential features (visible when joints are viewed edge-on) that are diagnostic of tension fractures (Kulander, et al., 1979).
   c. Transient features (visible when joints are viewed face-on) diagnostic of tension fractures (ibid).
2. True tensional stresses cannot persist below very shallow depths in the earth's crust, depths on the order of a few hundred feet, because of the effect of the lithostatic gradient.
3. Therefore, jointing is a near-surface phenomenon.

The fallacy in this line of reasoning lies in the assumption that only true tensional stresses can cause rocks to fail in tension.

Figure 13 depicts the composite failure envelope for brittle materials, including rocks, together with the Mohr's Circle representation of stress. If compressive stress is defined as positive with values increasing to the right, Mohr's circles representing possible stress states in the earth's crust (compressive stresses only) are confined to the right of the ordinate. The diagram shows that those circles large enough to contact the composite failure envelope do so only in the linear portion. The resulting ruptures are shear failures (i.e., faults); tension failures are impossible.

Rocks, however, do not fail in response to absolute stress; they fail in response to effective stress (Jaeger, 1962, p. 166; Hubbert and Rubey, 1959), which is related to absolute stress in the following manner:

$$\sigma = \sigma - p \qquad (1)$$

where:

σ = effective stress

σ = absolute stress

p = the pore pressure exerted by interstitial fluids.

In any geologic environment where the fluid pore pressure is greater than the absolute stress, this relationship states that the effective stress will have a negative sign and, thus, be tensile. The relevance of that conclusion to the formation of joints can be seen in figure 14, where the normal stress axis is relabeled to read effective stress ($\bar{\sigma}$) rather than absolute stress (σ). In this reference system, for a given absolute stress state remaining constant, increasing pore pressure has the effect of displacing the Mohr's circle to the left, eventually driving it against the failure envelope. If the circle, the diameter of which is equivalent to the maximum stress difference ($\bar{\sigma}_3 - \bar{\sigma}_1 = \sigma_3 - \sigma_1$), is sufficiently small, it will contact the envelope where it crosses the abscissa, to the left of the ordinate, and tensile failure will result. Inasmuch as the point of tangency occurs on the abscissa, the angle 2Θ between the compressional axis and the radius to that point will equal 180°, implying that the angle between the fracture plane and the least principal stress will be 90°. In simpler terms, figure 14 predicts that tensile fractures or joints can occur in rocks under conditions of elevated interstitial fluid pore pressure, provided the ambient stress anisotropy is relatively small; It further predicts that they are created in an orientation perpendicular to the direction of the minimum compressive stress. Thus, the mechanics of geologic jointing can be viewed as exactly analogous to that of artificial hydraulic fracturing (see Hubbert and Willis, 1957 for discussion on hydraulic fracturing mechanics). The most important consequence of this conclusion is the perception that jointing need not be a depth limited phenomenon.

This insight on the fundamental mechanics of jointing is not original with the authors. Donald T. Secor (1965) originally proposed it; though, it is not yet universally accepted. His exposition of what could be termed "the hydraulic fracturing theory of jointing" is more highly developed than the foregoing. In figure 15, an adapted version of Secor's (1965) figure 3, the normal stress axis (abscissa) on both sides of the shear stress axis (ordinate) has been graduated in multiples of "-K," the tensile strength of any rock under consideration. The composite failure envelope crosses the normal stress axis at -K. Hence, from figure 14, the condition for jointing is:

$$\bar{\sigma}_3 = -K \qquad (2)$$

where:

$\bar{\sigma}_3 =$ the least effective principal stress

and, as before,

$-K =$ the tensile strength of the rock.

Figure 15 shows that when displaced to the left by elevated fluid pore pressure, all Mohr's circles smaller than a certain critical diameter are capable of initially contacting the composite failure envelope at -K, where it crosses the abscissa. Most larger circles initially contact the failure envelope to the right of the ordinate, with the result that faulting, rather than jointing, occurs. The critical circle is centered at +K and crosses the abscissa at +3K, as well as at -K. Inasmuch as that circle has a diameter of 4K, the criterion for jointing may be stated as follows:

jointing occurs when

$$\bar{\sigma}_3 = \sigma_3 - p < -K \qquad (3)$$

or equivalently

$$p - \sigma_3 > K \qquad (4)$$

provided that

$$\bar{\sigma}_1 - \bar{\sigma}_3 = \sigma_1 - \sigma_3 < 4K. \qquad (5)$$

Or, alternatively stated, jointing occurs whenever the interstitial fluid pore pressure exceeds the minimum principal compressive stress (absolute) by the amount of the tensile strength, provided the maximum stress difference does not exceed four times the tensile strength.

What evidence is there to support the contention that the natural fractures responsible for most gas production from the Devonian shale of the Appalachian foreland are joints created in the manner described above? Natural fractures observed in Devonian shale core fall into one of three categories: slickensided, mineralized, and unfilled. Slickensided fractures are clearly not joints, but faults, having experienced lateral displacement at some point in time. High-angle slickensided fractures are rare. Low-angle slickensided fractures are rare to nonexistent throughout much of the Appalachian Plateau, but they become common in the folded plateau and increase dramatically toward the Allegheny Front. They qualitatively reflect the degree of detachment activity in the Devonian shale. The role of slickensided fractures in Devonian shale production is as yet uncertain and the subject of some debate. Although some of the slickensided fractures are also mineralized, most of the mineralized natural fractures observed in Devonian shale core show no evidence of lateral displacement, i.e., no trace of slickensiding and no offset of bedding or pre-existing fractures. Faint radial and crescentic markings, characteristic of joints, however, are often discernible on the planar faces of both the unfilled and the mineralized fractures. These are the transient fractographic features described and discussed at length by Kulander, Barton, and Dean (1979). These fracture face markings are diagnostic of **extensile** fracturing, i.e., jointing.

That most of the joints in Devonian shale **core** are mineralized (primarily calcite and/or dolomite), at least to some degree, is mute testimony to the presence of aqueous fluids at some point in their geologic history. The joints cannot be true tension fractures resulting from erosional unloading, because most of the EGSP Devonian shale cores were extracted from depths of several thousand feet, depths that are an order of magnitude greater than those at which true tensile stresses can exist in the earth's crust. Moreover, since deposition, each Devonian shale unit has been subject to nearly continuous and often rapid burial until the onset of the modern cycle of erosion initiated by the Alleghenian Orogeny. **Also,** the introduction of mineralizing fluids into these joints is hardly unrelated to their creation, since:

1. The walls of mineralized fractures in the Devonian shale are commonly separated by the mineral filling to a noticeable degree, 1 to 3 mm or more, implying that the mineralizing fluid was under pressure, forcing the fracture open while mineral matter was precipitating on the walls.
2. In certain instances the mineralization has a noticeably linear fabric, a semi-fibrous appearance, that is oriented perpendicular to the plane of the fracture, implying that distension of the fracture and mineralization occurred concurrently.

Grounds exist for inferring that the fractures were created and mineralized after the lithification of the Devonian shale, but prior to the establishment of the modern geologic environment. They cannot be penecontemporaneous with deposition, because they are true brittle fractures, as indicated by their generally planar habit and the diagnostic face markings. To behave as a brittle substance, the shale had to be at least well consolidated, if not indurated. On the other hand, no detectable mineralizing fluids have yet been encountered during coring operations, which accords with the reputation the Devonian shale has among drillers and producers of being a water-free formation. Moreover, most mineralized fractures are now so completely so, as to be virtually impermeable to aqueous fluids (though presumably not to gas). Thus, most of the jointing, or extension fracturing, in the Devonian shale probably occurred during the waning stages of dewatering, but not until induration was well advanced.

CONCLUSION AND SYNTHESIS

The Devonian shale of the eastern interior basins is a potentially major resource of natural gas, largely unexploited because of its exceedingly low permeability and consequently low average per well production rates. The weight of evidence, acquired under or made available through the Department of Energy's (DOE) Eastern Gas Shales Project impels the authors to conclude that natural fracture **systems** are indispensable to Devonian shale production, even with modern well stimulation technology. The natural fractures to which production can be attributed appear for the most part to be joints rather than faults, i.e., extension fractures rather than lateral displacement fractures. With the recognition that rocks fail under effective stress rather than absolute stress comes the realization that there is theoretically no limit to the depth at which jointing can occur, provided the internal fluid pore pressure exceeds the minimum principal stress (compressive) by the amount of the tensile strength of the rock and the maximum principal stress difference is not greater than four times that amount. Carbonate mineralization typical of extension fractures observed in Devonian shale core bears mute testimony to the role of pressurized fluids in their creation. The authors, therefore, conclude that the natural fractures essential to production from Devonian shale are dominantly joints that originated as natural hydraulic fractures, induced by excessive fluid pore pressure that built up at various times during its post-depositional history.

These natural hydraulic fractures, like their artificial counterparts, must have aligned themselves parallel to the plane of the two then prevailing maximum principal stresses, i.e., perpendicular to the minimum principal stress. Thus, joints in the Devonian shale represent a cumulative, though fragmentary, record of its stress history. This hypothesis suggests an exploration strategy for fractured reservoirs in the shales. Were it exposed at the surface, it should **be** possible to unravel some of the tectonic history of the Devonian shale by establishing a relative chronology among joint sets (see Kulander, et al., 1979). For the shale in the subsurface, perhaps the process can be profitably reversed. Using the tectonic histories of marginal areas and the underlying basement as boundary conditions, the exploratlonist should try to synthesize a tectonic history for the Devonian shale and use it to predict potentially resulting joint patterns, from which he could identify likely areas of concentrated fracturing.

Of particular concern to the Devonian shale explorationist in the Appalachian Basin are possible interactions between the zones of influence of cover-restricted detachment (thin-skinned) tectonics and basement tectonics. Overbey (1976) was intrigued by this possibility following his discovery that the near-surface, maximum horizontal in situ principal stress, which seems to have a regional east-west orientation throughout much of the Appalachian Basin, was reoriented over the Rome Trough, a buried rift system, to conform to its strike.

The authors formulated two supposedly competing hypotheses to explain this phenomenon. The basis for each was some postulated interaction between thin-skinned and basement tectonics. Both hypotheses presume the existence of buried normal faults that markedly offset the basement unconformity in relationship to the total thickness of the sedimentary cover (see Harris, 1975 and 1978, also Kulander and Dean, 1978). They differ in the proposed manner by which the buried faults interact with the detachment tectonics characteristic of the Appalachians. The'buried fault buttress model"(figure 16) postulates a passive basement with a pre-existing normal fault and an actively sliding cover. The"buried fault slip model"(figure 17) postulates a passive cover under the influence of detachment related lateral stress and a rejuvenated normal basement fault slipping a small fraction of the amount of the total pre-existing displacement.

To evaluate the plausibility of the two hypotheses the EGSP let a small contract to West Virginia University (WVU) for Professors S. H. Advani and H. V. GangaRao of the Engineering and Applied Science Department and their student to conduct finite element stress analysis on the two idealized models (figures 16 and 17). Involving a high-speed computer, finite element stress analysis is a means of solving for the complete state of stress (intensity and orientation) at any point within a continuous medium, given boundary conditions and physical properties. The results of this investigation, which in the larger sense was also a feasibility study on the application of finite element stress analysis to tectonic problems, were summarized in a paper presented at the First Eastern Gas Shales Symposium (Advani, GangaRao, Chang, Dean, and Overbey, 1977). From a strictly theoretical viewpoint, both of the models turned out to be equally plausible explanations for the in situ stress reorientation observed at the surface. The output of the NASTRAN computer program, which performed the analysis, enabled S. H. Advani and company to contour stress intensity in the cross-sectional view of the two models, figures 16 and 17. The horizontal stress intensity contour plots for the two models both displayed a near-surface zone of relative tension over the buried basement fault. Had the models been formulated in three dimensions, this condition would have been manifested as a reversal of the relative magnitudes of the two horizontal principal stresses, which is equivalent to the in situ stress reorientation observed in nature.

Though the two models predict indistinguishably similar surface stress configurations, the predicted subsurface configurations differ very significantly. Stress intensity plots for the buried fault slip model (figure 17) reveal subsurface zones of low stress and even relative tension. Were conditions corresponding to those of the buried fault slip model to have occurred in nature, the equivalent zones would have been ideal sites for concentrated jointing, i.e., natural extensional fracturing by the mechanism described in the preceding section.

The question yet remains, which of the two proposed hypotheses idealized in the cross-sectional models (figures 16 and 17) best explains the observed phenomenon of reoriented stress over the Rome Trough? This and similar questions are not academic, as correct answers may point the way to subsurface fractured reservoirs in the Devonian shale. Sophisticated exploration techniques, to wit, finite element stress analysis, can theoretically identify environments particularly prone to natural fracturing within a specified tectonic context. The problem lies in specifying the tectonic context, which corresponds to the boundary conditions of the model to be analyzed. The Devonian shale explorationist is limited by his ignorance of tectonic influences upon the shale. Clarifying some of these uncertainties is the raison d'être of this volume, "Western Limits of Detachment and Related Structures in the Appalachian Foreland".

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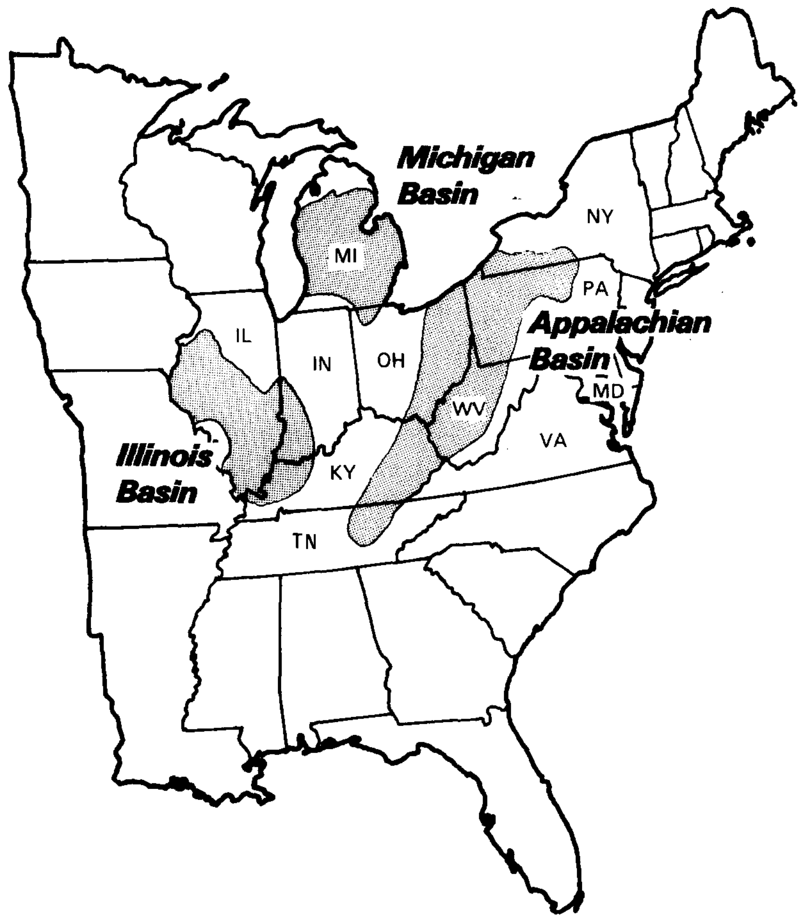
1: present address: Arco Oil and Gas Co., P.O. Box 5540, Denver, Colorado 80217.

Known Gas Shale Deposits Located in the United States



Eastern Gas Shale Deposits

Figure 1. Distribution of Devonian shale in the eastern United States.

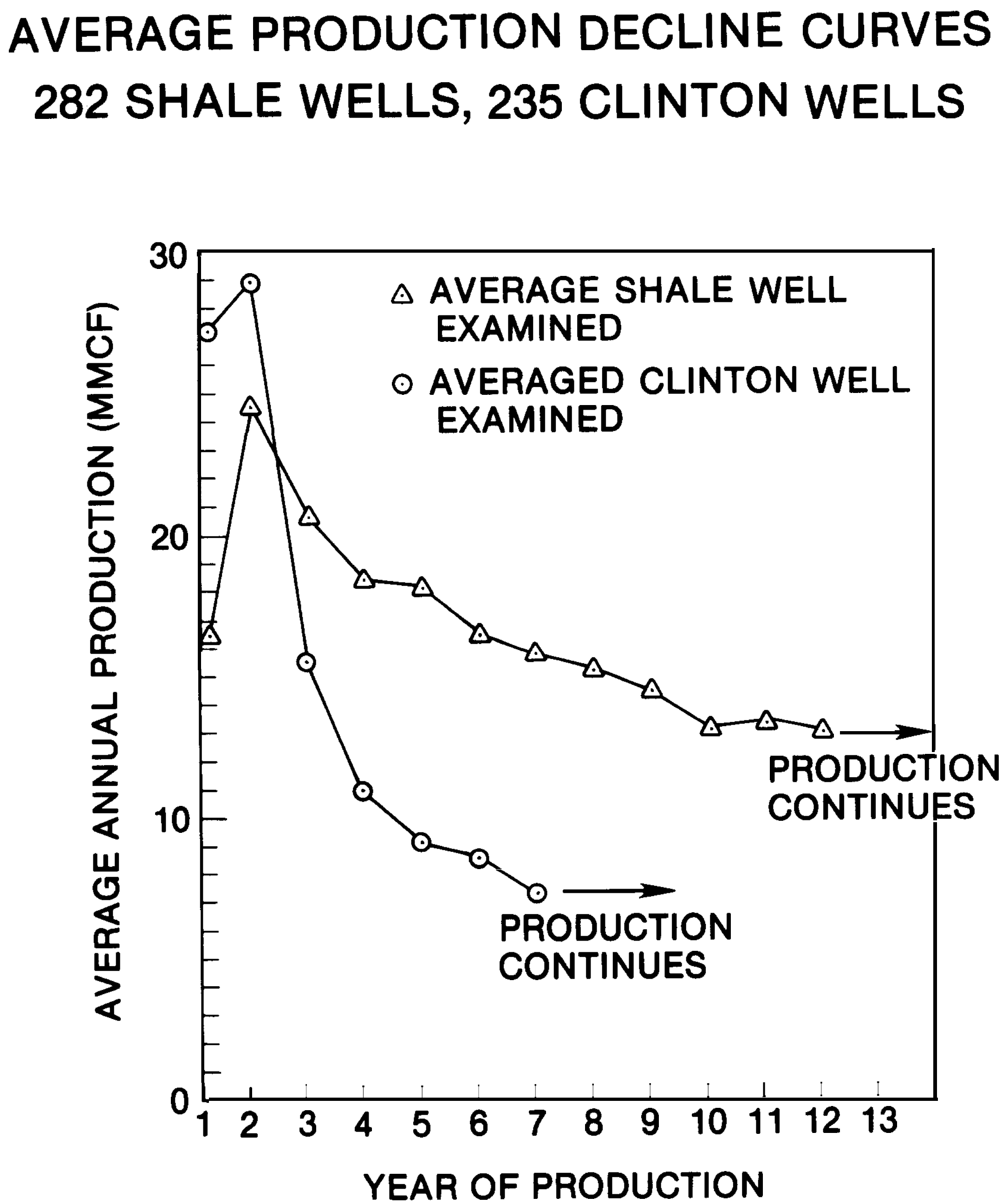


Figure 2. Average production decline curves for unconventional shale wells compared with conventional sandstone wells (Brooks, Forrest, and Morse, 1974, figure 5).

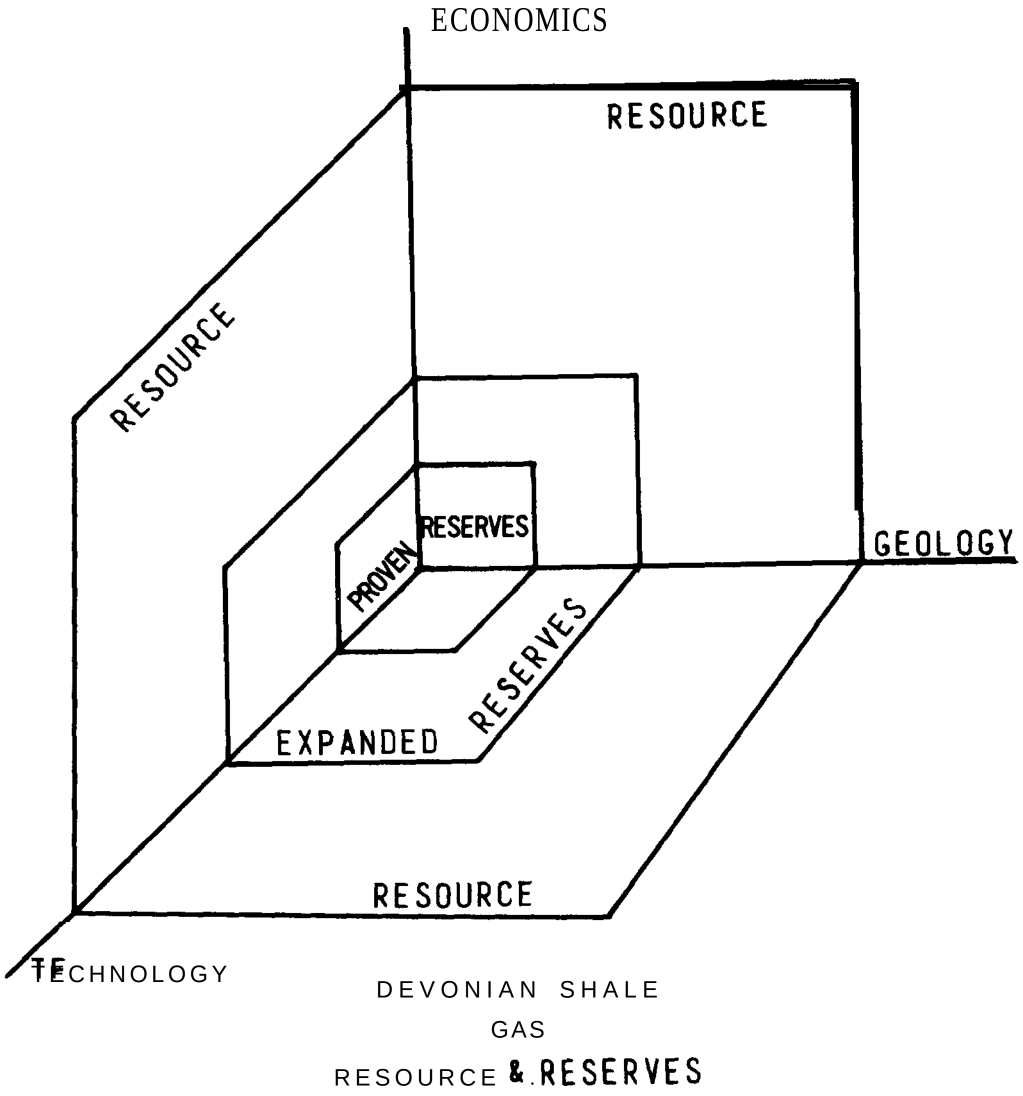


Figure 3. Dependency of Devonian shale gas reserves upon economics, technology, and geology.

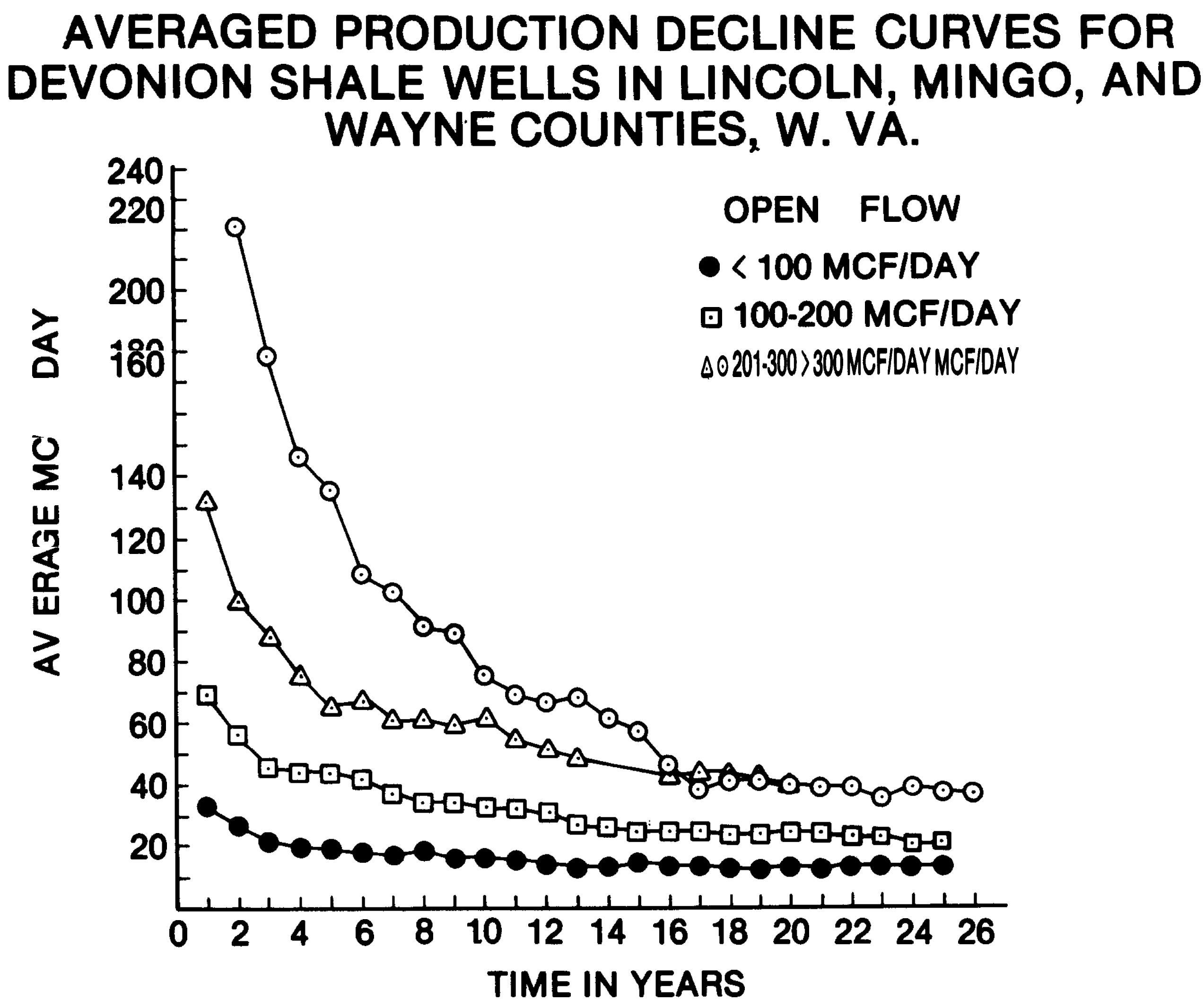


Figure 4. Averaged production decline curves for Devonian shale wells differentiated according to open flow (Bagnall and Ryan, 1976, figure 11).

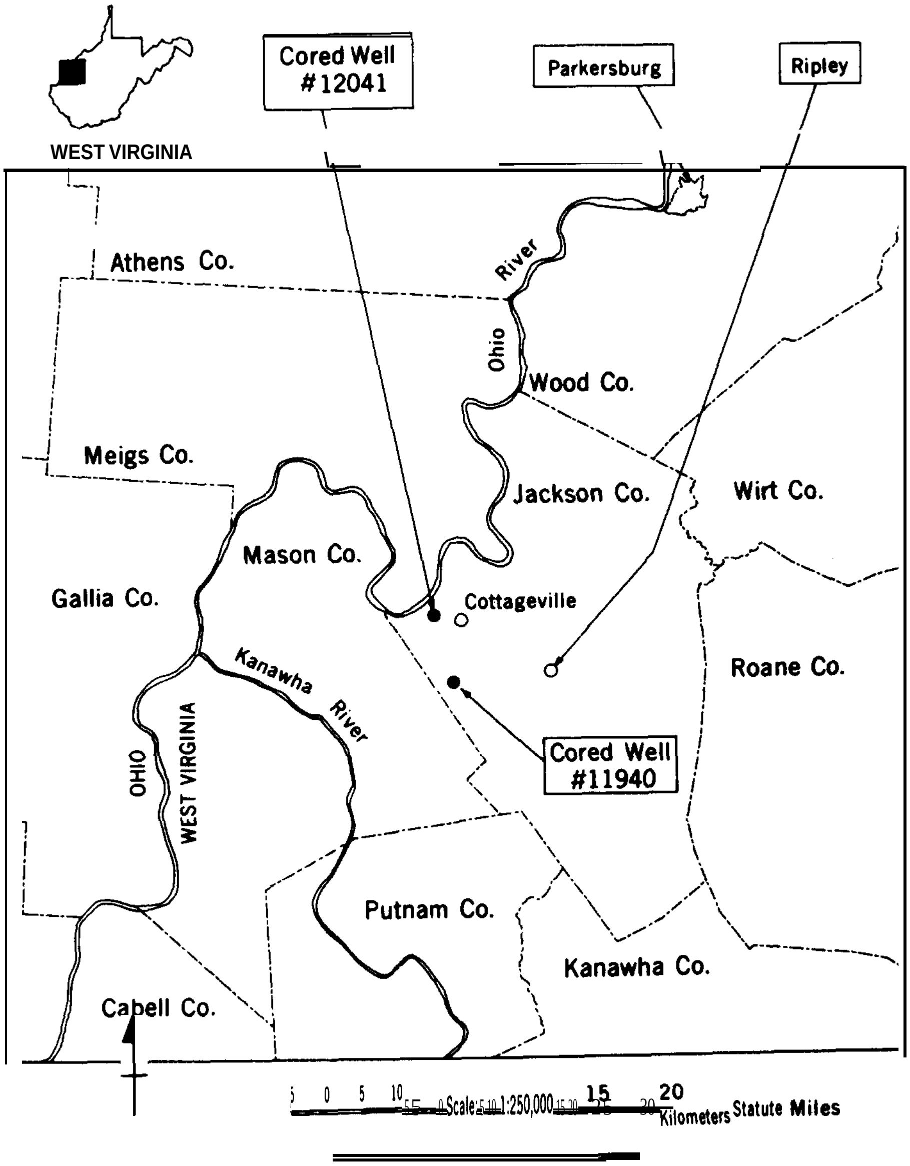


Figure 5. Location of cored Devonian shale wells in the Cottageville gas field of Jackson and Mason Counties, West Virginia (Byrer, Trumbo, and Rhoades, 1976, figure 1).

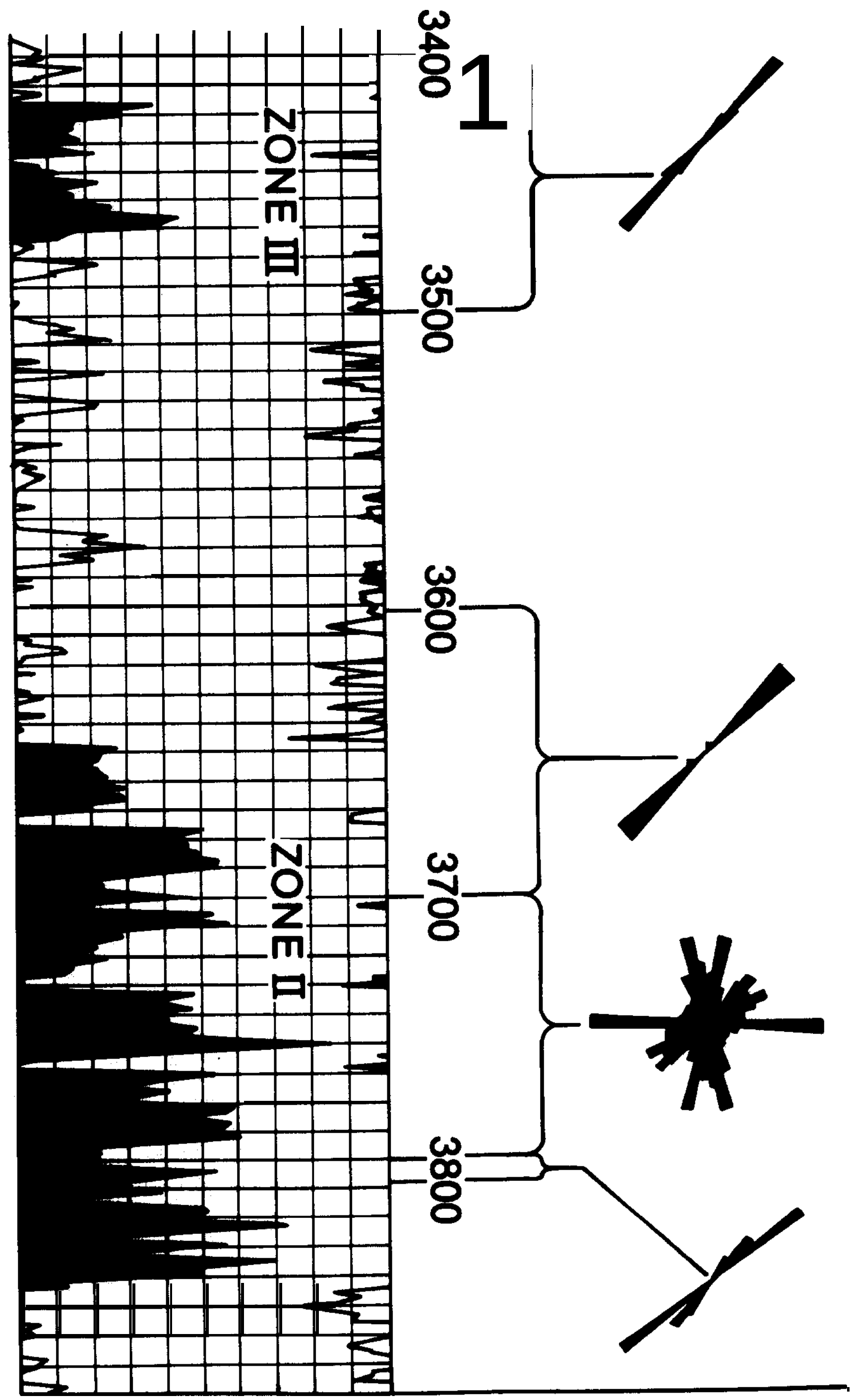


Figure 6. Gamma ray log and core fracture orientations from CGSC 11940, Jackson County, West Virginia (after Martin and Nuckols, 1976, figure 13).

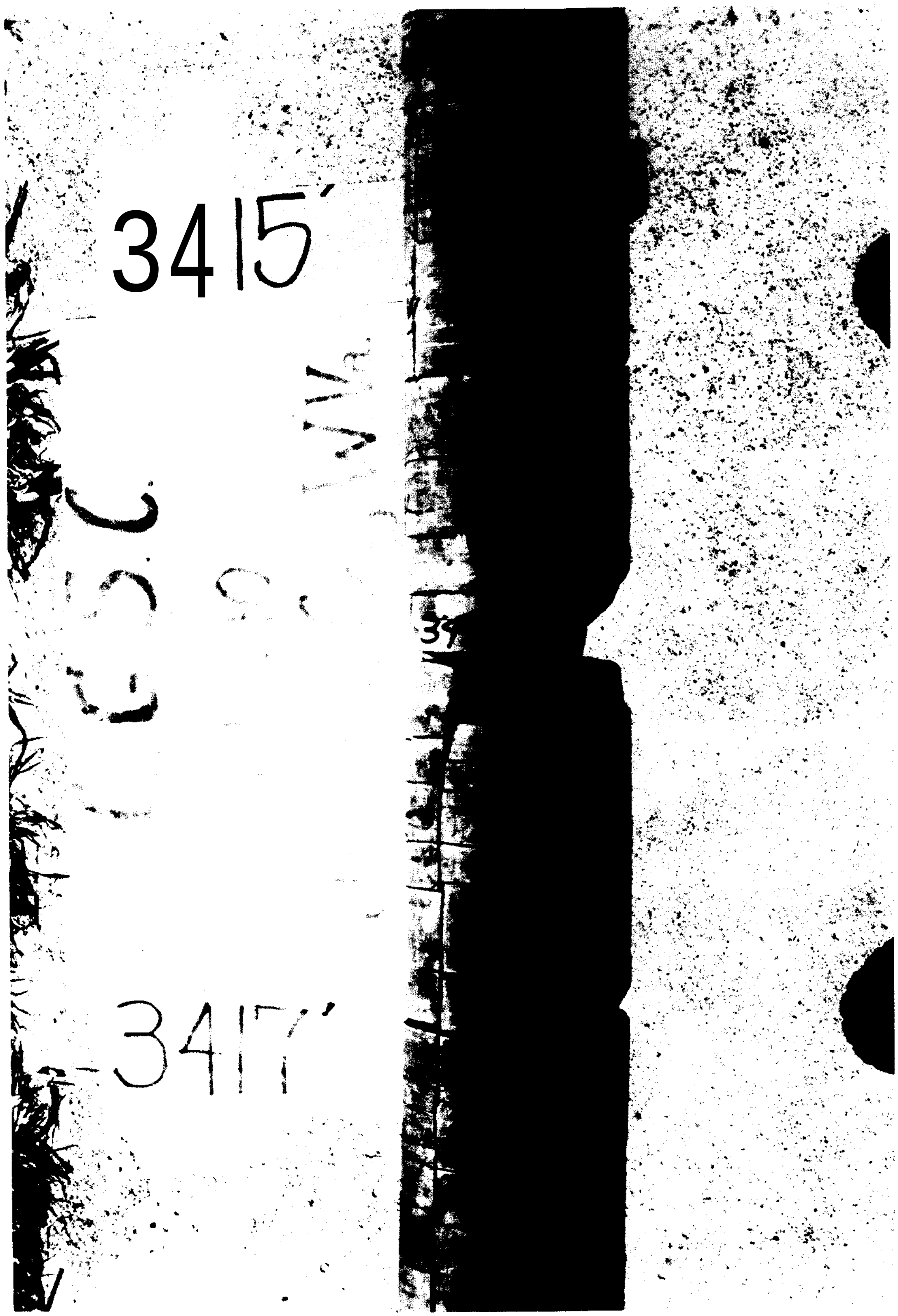


Figure 7. Edge-on view of a petal centerline and two petal fractures in the oriented core extracted from CGSC 11940, Jackson County, West Virginia.

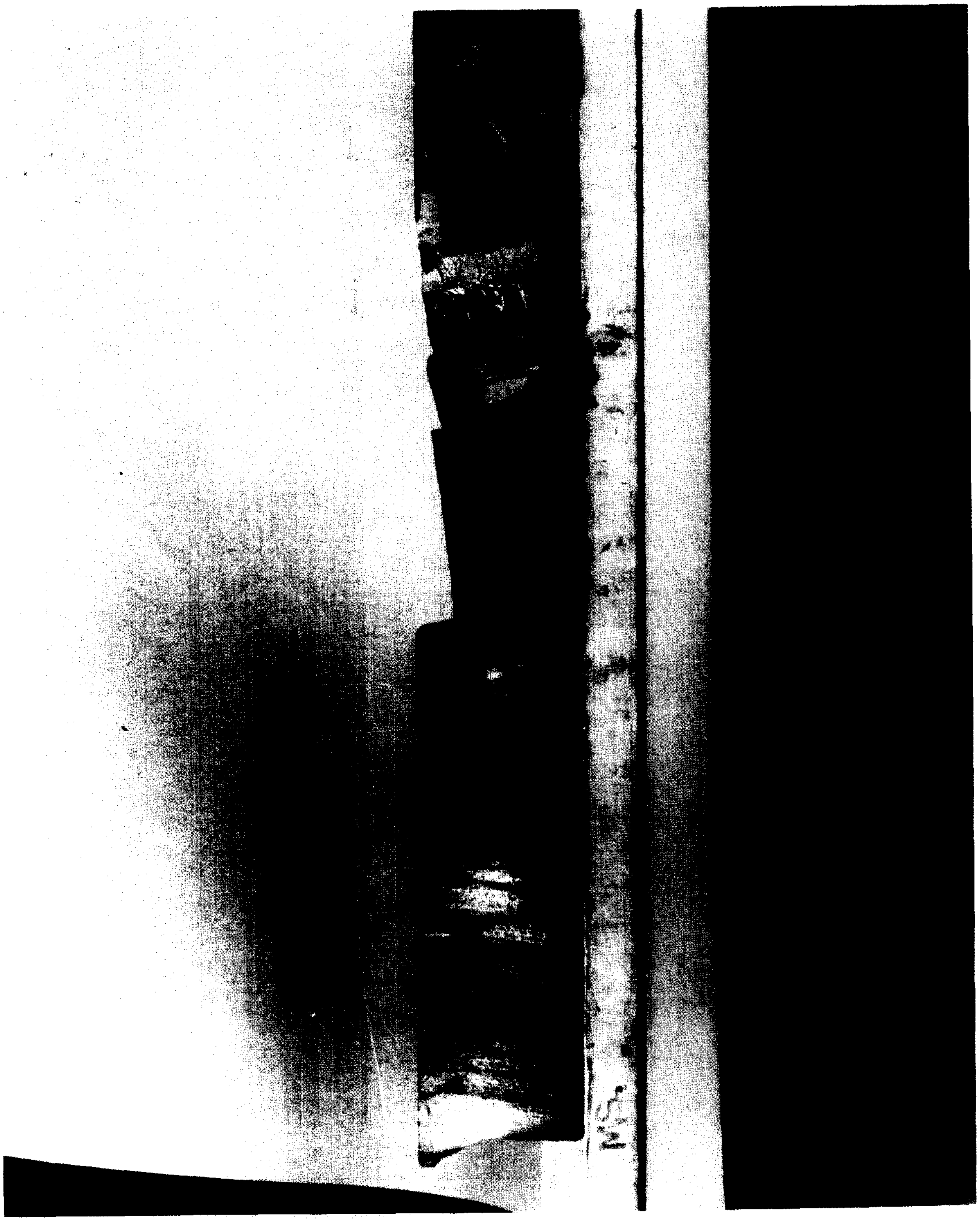


Figure 8. Fractures, coring induced and natural, in the CGSC 11940 core. The left hand portion (bottom of page) reveals a face-on view of a centerline fracture. Note the closely spaced intermediate arrest lines, indicating that the fracture propagated down-core by discrete extensions.

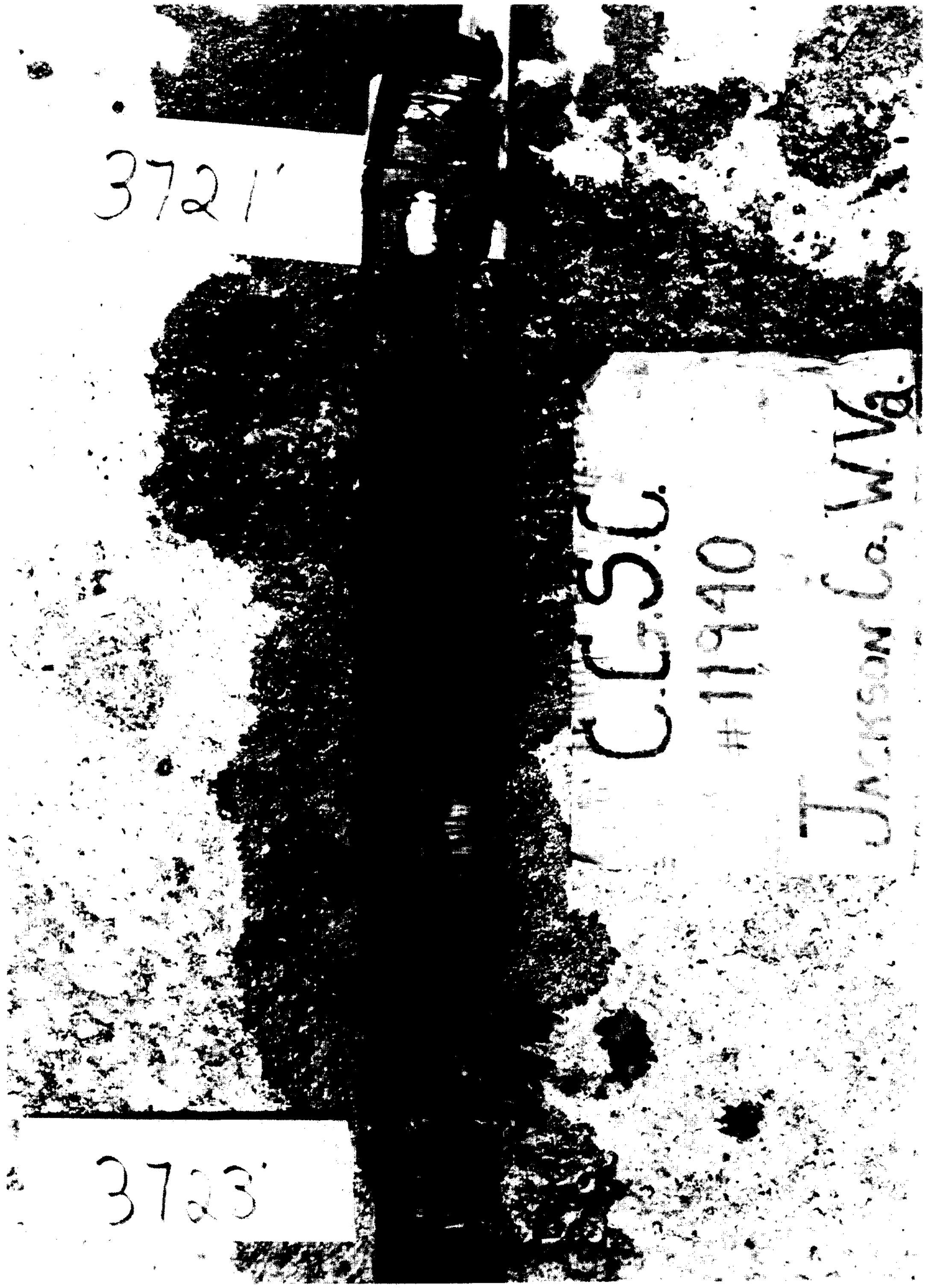


Figure 9. Diversely oriented, intersecting natural fractures in the CGSC 11940 core.

Figure 10. Close-up of intersecting natural fractures in the CGSC 11940 core.

Figure 11. Heavy dolomite mineralization on the face of a natural fracture in the CGSC 11 940 core.

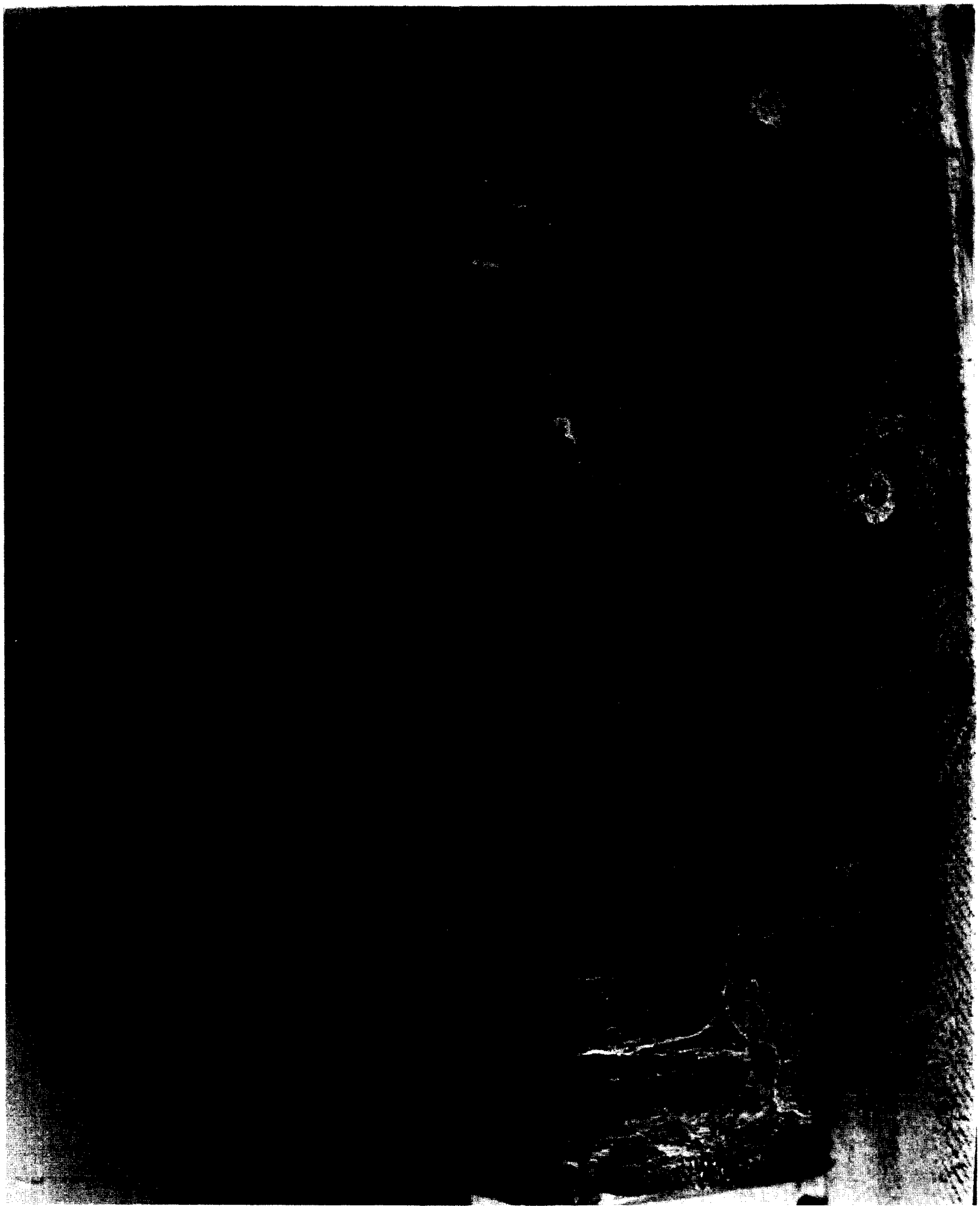


Figure 12. Dolomite mineralization on the faces of two intersecting fractures in the CGSC 11940 core.

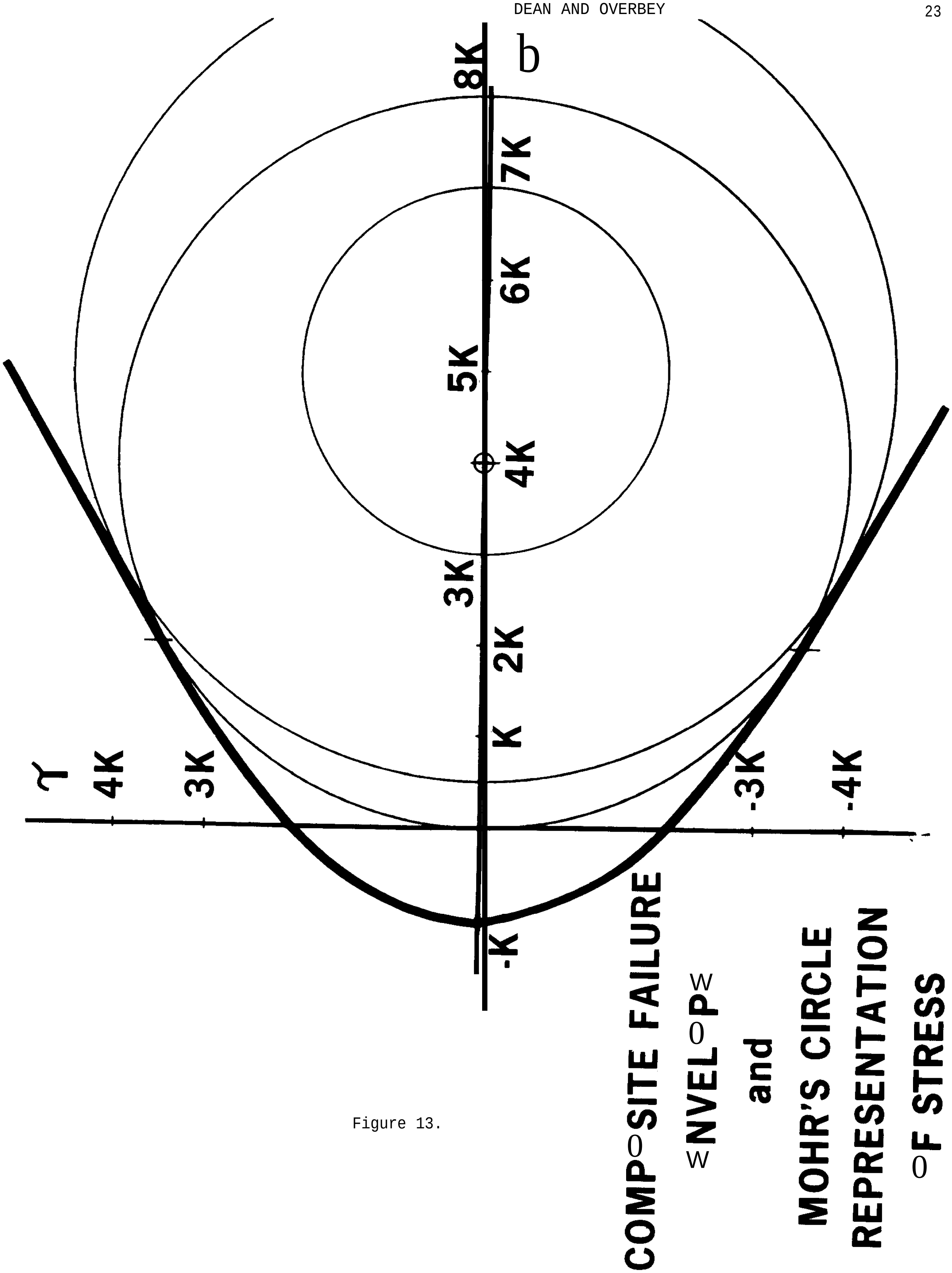

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Figure 13.

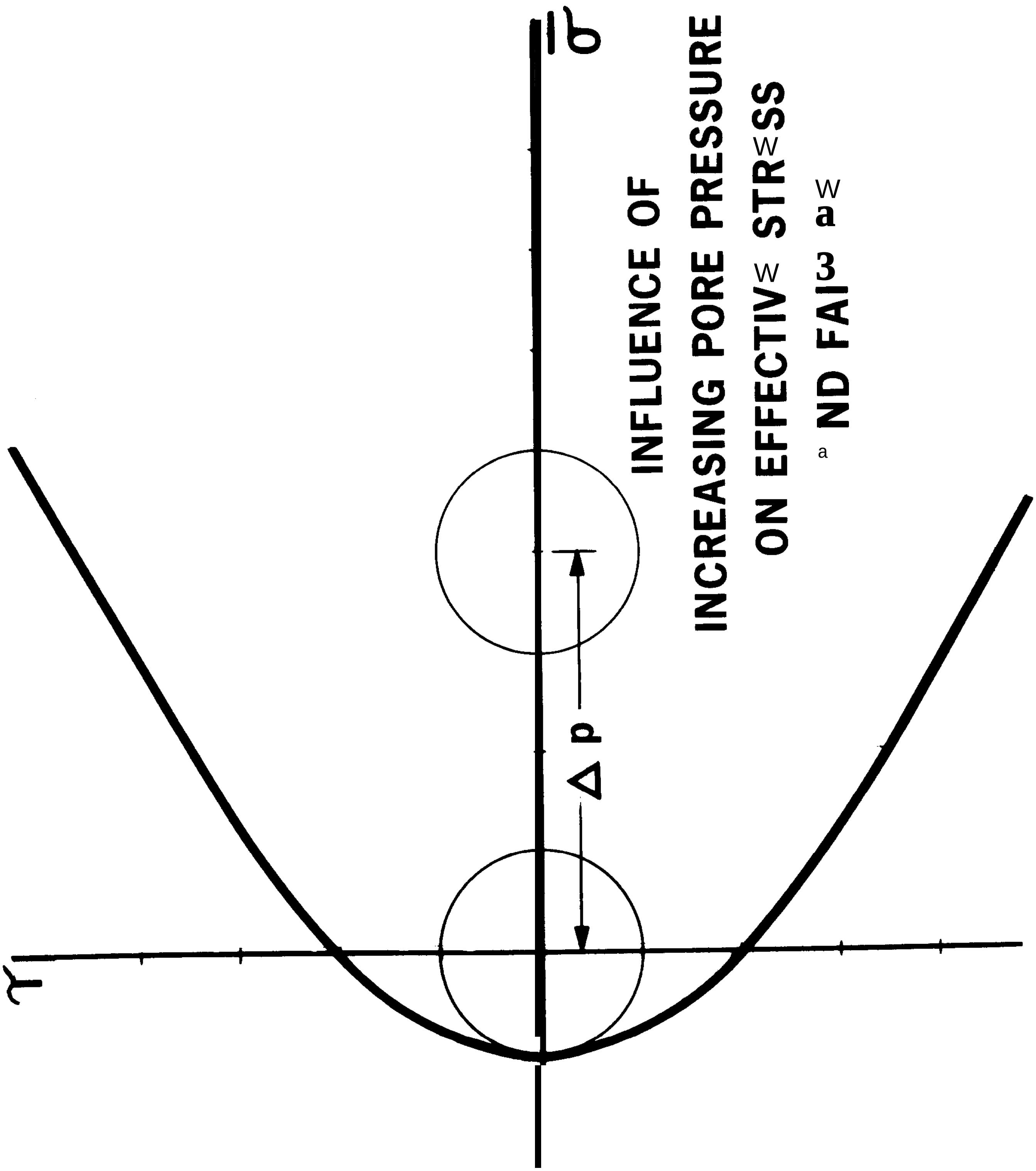

Figure 14. The influence of fluid pore pressure on the Mohr's Circle representation of effective stress. Note that increasing pore pressure displaces the circle progressively to the left, eventually driving it against the failure envelope.

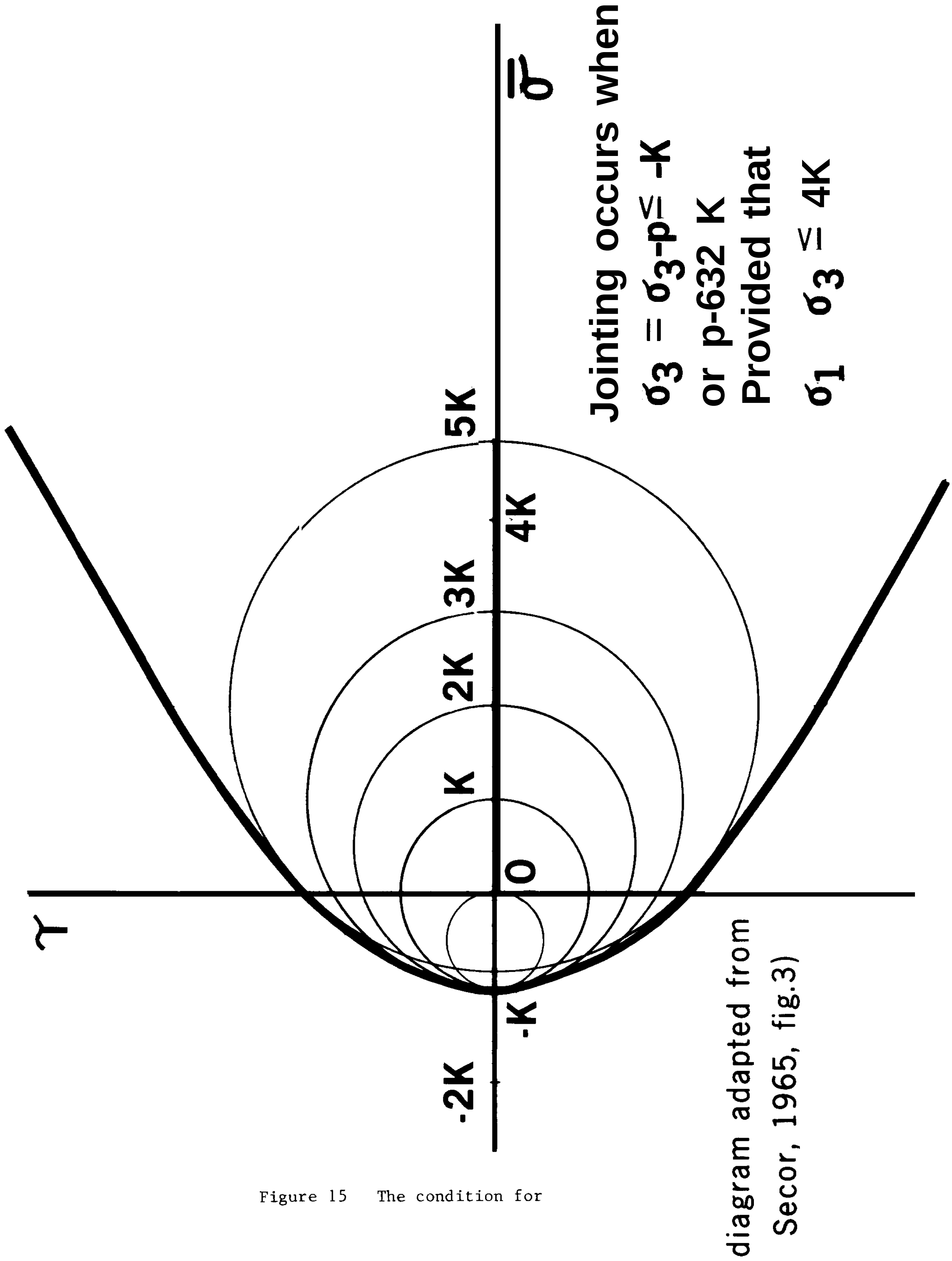

Figure 15 The condition for

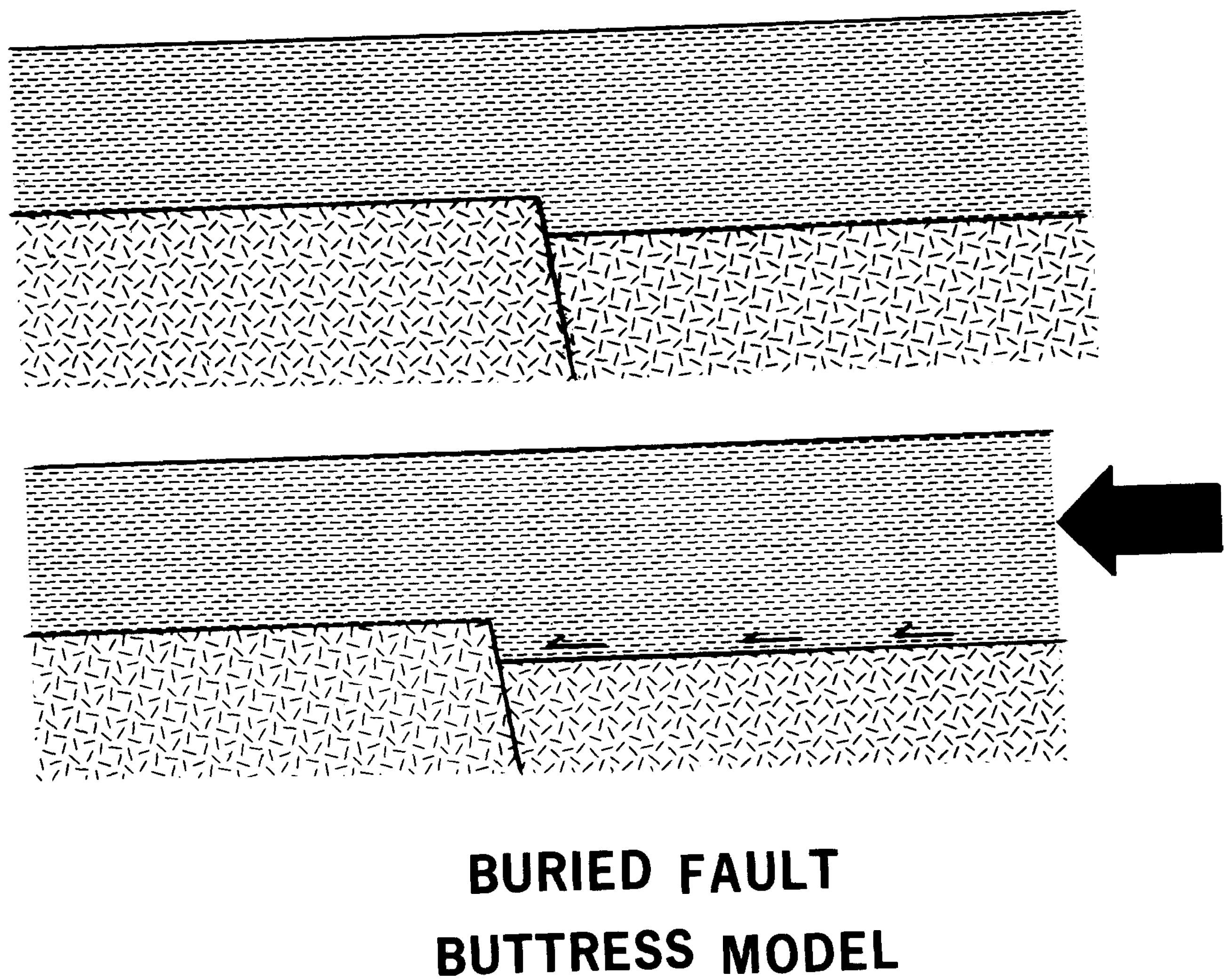

Figure 16. The "buried fault buttress model", the first of two proposed to explain the apparent reorientation of the near surface in situ stress over the Rome Trough. The model postulates a passive basement with a pre-existing normal fault and an active cover capable of sliding at or near the basement unconformity.

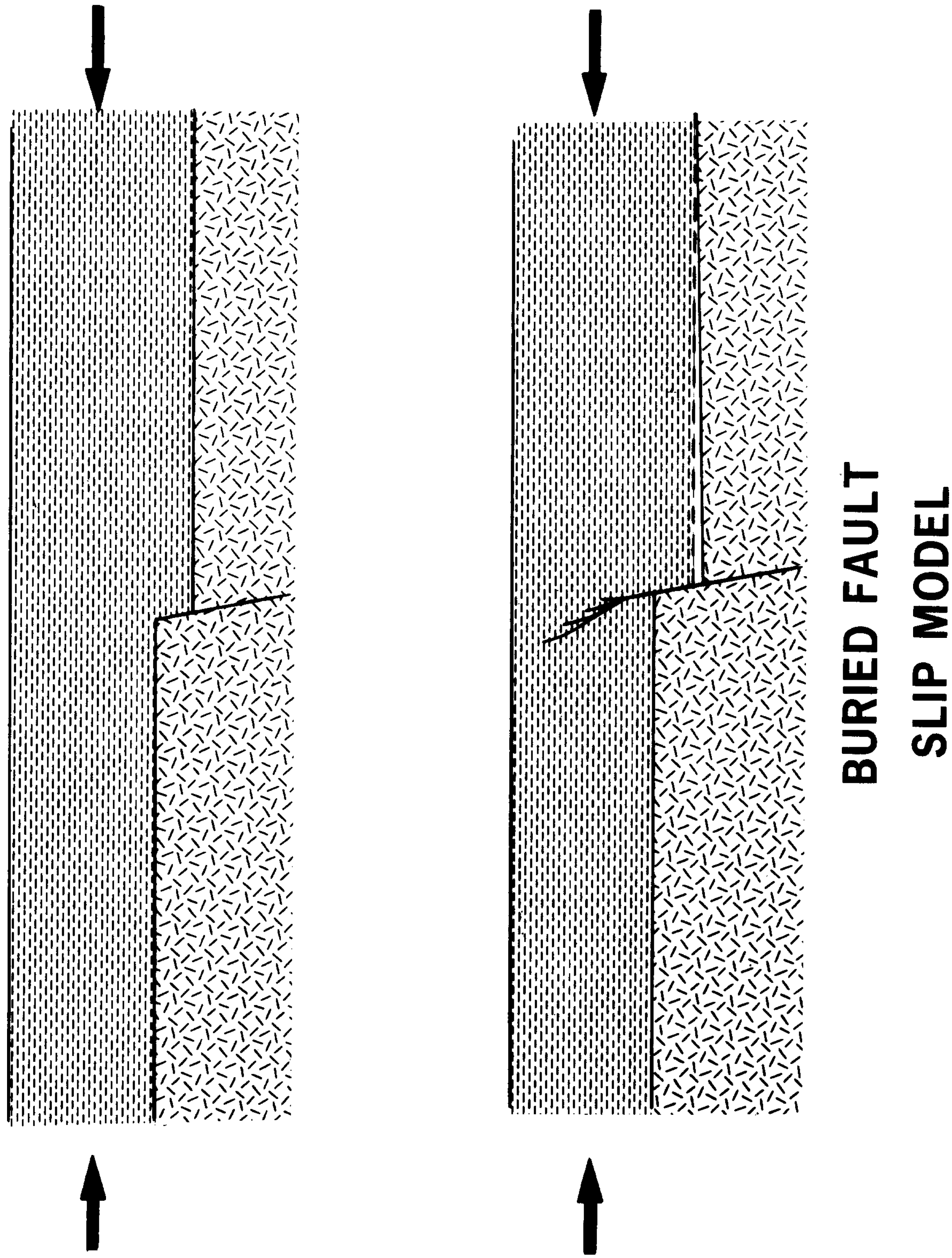

Figure 17. The "buried fault slip model", the second of two proposed to explain the apparent reorientation of the near-surface in situ stress over the Rome Trough. The model postulates a passive cover under the influence of detachment related lateral stress and a rejuvenated normal basement fault slipping a small fraction of the amount of the total pre-existing displacement.

STRATIGRAPHIC DISTRIBUTION OF DECOLLEMENTS IN ME APPALACHIAN BASIN[1]

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ABSTRACT

Décollement zones in the Appalachian basin occur in strata ranging in age from Cambrian to Pennsylvanian. These décollement surfaces follow shale and evaporite strata, bentonites, and unconformities. Widespread stratigraphic units control the distribution of major dbcollements affecting broad areas of the Appalachian basin. Some stratigraphic zones of more local extent form minor décollement horizons.

Ddcollement surfaces are not well understood for the Taconic deformation belt, but appear to be totally restricted to Ordovician Shale in Pennsylvania, New Jersey, and New York. Some minor décollements may have originated during the Acadian orogeny. The majority of dicollements formed during the Alleghany orogeny.

Maps were made to portray the geographic distribution of dicollement surfaces known in the Rome Formation, Sevier and Athens Shale, bentonites of the Moccasin Formation, Salina evaporites, Mandata Shale, Tioga Bentonite, Devonian shales variously known as Marcellus, Millboro and Chattanooga Shale, Maccrady evaporites, Floyd Shale, the unconformity between the Pottsville Formation and the underlying Mauch Chunk-Pennington strata, and shales within the Pennsylvanian System. Examples of tectonic style are given for each of these décollements.

The stratigraphic climbing of décollement surfaces toward the west may reflect the character of the deforming forces, but it more likely reflects the progressive development through time of appropriate lithologies extending farther west onto the craton. Important development of décollements generally extends westward only to the present structural axis of the Appalachian basin.

1: reprinted with permission from Geological Society of America Abstracts with Programs, v. 10, no. 4, p. 167 (1978).

THE NORTHWESTERN EXTENT OF COMPRESSION DURING DÉCOLLEMENT TECTONICS OF THE APPALACHIAN PLATEAU [1]

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ABSTRACT

The orientation of a maximum layer parallel shortening during compression of the Appalachian fold and thrust belt is NNW to NW in central Pennsylvania. That such a compression was transmitted to the vicinity of Buffalo, New York, a distance of 225 km from the Valley and Ridge Province in Pennsylvania is shown by several different types of strain measurements. All indicate a maximum layer parallel shortening oriented about NNW in western New York. Residual strain was measured by *in situ* strain relaxation techniques and by x-ray diffraction techniques. Nonrecoverable strain was indicated by deformed fossils, calcite twinning and solution cleavage. This penetrative strain within the Appalachian Plateau must be accounted for in palinspastic reconstructions of the Appalachian Mountains.

1: reprinted with permission from Geological Society of America Abstracts with Programs, v. 10, no. 4, p. 168 (1978).

For details see:

Engelder, T., 1979, Mechanisms for strain within the Upper Devonian clastic sequence of the Appalachian Plateau, western New York: American Journal of Science, v. 279, p. 527-542.

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EVIDENCE FOR MULTIPLE STRATIGRAPHIC STRUCTURAL HORIZONS IN THE VALLEY AND RIDGE AND APPALACHIAN PLATEAU OF THE CENTRAL APPALACHIANS [1]

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ABSTRACT

Recent studies in the Valley and Ridge and Appalachian Plateau of Pennsylvania, New York, and Maryland have revealed several sets of regionally developed minor structures (solution cleavage, wedging, fossil distortion), each of which appears related to particular stratigraphic horizons. Finite strain analysis and dynamic considerations suggest that each set is related to the development of individual decollements. These sets of structures, which are believed to differ in age of formation from place to place, have been termed stratigraphic structural horizons. Evidence supporting this concept in terms of shallow (50 -500 m) depth of formation and association with individual decollements is assessed in light of studies in the Umbrian Apennines and the Appalachian Plateau of eastern and central New York.

1: reprinted with permission from Geological Society of America Abstracts with Programs, v. 10, no. 4, p. 169 (1978).

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MAJOR CROSS STRIKE STRUCTURES IN PENNSYLVANIA[1]

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ABSTRACT

Cross-strike topographic features in the Appalachian Fold Belt may represent (a) antecedent valleys from an inherited drainage regime, (b) fault valleys, or (c) fracture controlled streams. Satellite imagery facilitated their mapping, but paucity of outcrops and large size has. hampered bedrock characterization. Preparation of the new State Map has focussed attention on these features on a scale appropriate to their size.

Recent mapping has shown the lineaments to be zones of structural disturbance, characterized by thrust and tear fault combinations of complex form. In the Blue Ridge section of the Bedford-Everett 1 ineament, the Carbaugh-Marsh Creek fault changes from a steeply dipping fault with right lateral separation to a thrust fault. Similar complex tear and thrust faults occur in the Everett Gap - Friends Cove faults, and the Bedford - Wells Mountain faults farther west. In between are the Breezewood and Sideling Hill faults, with apparent right lateral and vertical motions in different places. Magnetic and seismic anomalies to the west imply a structural disturbance well into the Allegheny Plateau. The McAlevys Fort - Port Matilda lineament is a scaled down version of tear and thrust. faults. The Ty rone - Mount Union 1 ineament appears to be a decoupled zone described by increased fracture density, and tilt-meters have been installed to detect any displacements due to earth tides.

Some lineaments are expressed in the gravity and seismic maps. Magnetic and seismic perturbations attest to a subsurface disturbed zone. Gas accumulations might be expected in finely fractured zones beneath a decollement or self-healing cap rock, e.g., in the Allegheny Plateau, and truncated anticlinal structures such as Hoskins & Root (1977)[2] reported north of Breezewood.

1: reprinted with permission from Geological Society of America Abstracts. with Programs, v. 10, no. 4, p. 170 (1978). See also Wheeler, this volume, and references cited there.

2: Root, S.I., and Hoskins, D.M., 1977, Lat. 40°N fault zone, Pennsylvania: a new interpretation: Geology, v. 5, p. 719-723.

WESTERN LIMIT OF EXTENSION FRACTURING IN WEST VIRGINIA

by

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ABSTRACT

Extension fractures can extend beds and create porosity, permeability or both where dips on anticlinal limbs are relatively steep (greater than 45 degrees if there is no strain parallel to hinge lines). Simple calculations and reasonable assumptions permit the conclusion that including the effect of hinge line parallel strain does not significantly alter the minimum bed dip at which extension fractures will open. Therefore the minimum conditions necessary for extension fracture formation are unlikely to occur west of Gwinn's (1964) high plateau, with the exception of the Burning Springs anticline, in Wood and Wirt Counties, West Virginia, where limb dips reach a maximum of 68 degrees.

INTRODUCTION

Detachment faults in ductile beds are the primary mechanism for map-scale deformation in the central Appalachian allochthon, and specifically for the Valley and Ridge and most of the Plateau provinces in West Virginia (Gwinn, 1964; Rodgers, 1963). The location of the western limit of detachment in the central Appalachians is a matter of conjecture, although it may be significant in explaining the distribution or occurrence of gas production from the Middle and Upper Devonian clastic sequence (mostly shales). The purpose of this paper is to determine the western limit in West Virginia of detachment-related fracture porosity and permeability formed as folding rotates beds from low limb dips, where they are contracted, to high dips, where they are extended.

STRUCTURAL STYLE

The West Virginia allochthon has many detached anticlines but few outcropping major thrust faults. The major anticlines are interpreted as active features generated mainly by duplication of strata by ramping of underlying detachment faults, by splay faults, by kink band folding, and by ductile flow of shale rich intervals into anticlinal crests (Gwinn, 1964; Faill, 1969; Wheeler, 1975). The major synclines are regarded as passive features resulting from anticlinal growth in adjacent rocks, rather than from active downbuckling (Gwinn, 1964).

Location of the western limit of allochthonous anticlines involving the Devonian clastic sequence is ambiguous, because the folds generally become less distinctive towards the west. Limb dips and closure decrease westward. In some of the folds of central and western West Virginia, such as the Arches Fork and Wolf Summit anticlines, closure is apparent below the Middle and Upper Devonian Gnesquethaw Group (Cardwell, 1973), but not above the clastic sequence, on the Middle Mississippian Greenbrier Group (Haught, 1968). The limestones, sandstones, cherts, and thin shales of the Onesquethaw Group and the underlying Oriskany Sandstone, Helderberg Group, and Tonoloway Formation commonly form a single stiff unit about 600 to 1000 feet (189 to 300 meters) thick (Cardwell and others, 1968) in the mechanical stratigraphy of West Virginia. If the folds in the Onesquethaw are part of the allochthon, then a detachment below the Silurian Tonoloway and the thick Devonian clastic sequence (2600 to 7800 feet; 800 to 2400 meters) has absorbed the deformation by flow (Cardea, 1956) or upper detachment (Dahlstrom, 1969). On the other hand, some folds show closure in the Mississippian Greenbrier Group but not in the Devonian Oriskany Sandstone, from which we infer that a detachment is above the Oriskany and probably in the ductile Middle Devonian black shales. Perry and Wilson (1977) describe an example of this on the Mann Mountain anticline.

We believe that extension fractures may form a significant part of gas-bearing fracture porosity and permeability. Extension fractures that show bed-parallel extension are interpreted to be late tectonic, formed when folding beds rotated to such steep dips that the rocks were within the extensional, rather than contractional field of the strain ellipsoid (Berger, Perry and Wheeler, ms. in review).[3] These extension fractures are unlikely to be filled by vein material or closed by later deformation and thus may produce gas. We attempt to locate the western limit of extension fracturing in the subsurface as part of a program to predict the location of more highly fractured rock for gas exploration.

In this paper we use the analytical and graphical techniques of Ramsay (1967) to determine the angle (θ) between the X strain axis (the direction of greatest lengthening: vertical) and the surface of no finite longitudinal strain. The surface of no finite longitudinal strain marks the boundary between the contractional and extensional fields in the strain ellipsoid and its dip increases with the horizontal contractional strain. Assuming that the maximum principal compressive stress and the Z strain axis (direction of greatest contraction) are horizontal and perpendicular to strike (Perry, 1971), then the minimum bed dip at which extension fractures can open is $90 - \theta$ degrees. Using published values of shortening or values measured on published cross-sections, we calculate the critical (minimum) bed dip at which extension fracturing will begin. Comparison of predicted values of critical dips with observed or estimated values of limb dips or shortening in anticlines allows us to predict the distribution of extension fracturing and related fracture porosity and permeability in West Virginia.

METHODS

Ramsay (1967, p. 128) determined the equation for the angle (θ) between the surface of no finite longitudinal strain and the X strain axis, assuming constant volume and no hinge line parallel strain ($e_2 = 0$)

$$\theta = \pm\cos^{-1}\sqrt{\frac{\lambda'_3 - 1}{\lambda'_3 - \lambda'_1}} \qquad (1)$$

where $\lambda'_i = 1/\lambda_i$

and $\lambda_i = (1 + e_i)^2, \qquad i = 1,2,3 \qquad (2)$

Therefore to determine the minimum bed-dip at which extension fractures will begin to open for a given anticline and its estimated shortening value, we need values of λ_1 and λ_3

λ_3 measures length change parallel to the X (contractional) strain axis. λ_2 measures length change parallel to the hinge line of the fold (intermediate strain axis: Y) and is assumed to be equal to 1. λ_1 measures strain parallel to the Z (extensional) strain axis. λ_3 is found by measuring the amount of shortening for a given anticline by the sinuous bed or equal area method, after removal of synfolding or postfolding slip on faults that changed bed length but did not rotate beds. We shall regard folding of a stiff layer within a volume of softer rock as grossly approximating overall homogenous nonrotational strain. We assume that the volume of rock containing the rotating stiff bed is not cut by through-going detachments and thus has not been deformed by shear on or near the detachment. That assumption is reasonable within a single anticline (Kulander, oral communication, 1978). Then by the sinuous bed method after restoration of fault slip, the shortening strain is

$$e_3 = (\ell - \ell_0) / \ell_0 \qquad (3)$$

where ℓ_0 = the arc length of a stiff unit like the Onesquethaw-Tonoloway sequence, and

ℓ = the linear distance between the two inflection points of the anticline (the distance into which ℓ_0 has shortened).

The sinuous bed method is valid for stiff units that have buckled, kinked, or faulted, but have not flowed internally or pressure-dissolved. The values of shortening in this paper are determined by the sinuous bed method but are consistent with values determined by the equal area method (Gwinn, 1970), which does not assume that there has been no internal flowage or mass-removing pressure solution.

Our values of shortening need not include shortening by layer parallel penetrative strain (Engelder and Engelder, 1977) or by formation of pressure solution cleavage (Geiser, 1977), because most of this shortening appears to have formed early, prior to folding (Geiser, 1970, Nickelsen, 1976, 1978; Berger, Perry and Wheeler ms. in review).[3]

λ_1 can be calculated as follows. For a unit sphere deforming at constant volume to a strain ellipsoid

$$V\ (\text{sphere}) = {}^4/_3\pi r^3 = {}^4/_3\pi$$
$$V\ (\text{ellipse}) = {}^4/_3(1+e_1)(1+e_2)(1+e_3)\pi$$

Substituting equation (2) and setting V (sphere) equal to V (ellipsoid)

$$\lambda_1 = {}^1/\lambda_2\lambda_3 \qquad (4)$$

Therefore to determine λ_1 we need only λ_3, which we estimate from shortening, and λ_2, which we assume is equal to 1. Therefore

$$\theta = \pm\cos^{-1}\sqrt{\frac{1/(1+\ell-\ell_0/\ell_0)^2 - 1}{1/(1+\ell-\ell_0/\ell_0)^2 - (1+\ell-\ell_0/\ell_0)^2}} \qquad (5)$$

If we take into account hinge line parallel strain, then λ_2 will be close to but not equal to 1. We assume that e_2 will be no more than $\pm$10 percent of e_3. This figure is arbitrary but is probably in excess of the true value in most places, especially where large folds are straight in map view. There is no doubt that strain parallel to hinge lines occurs. The existence of cross joints in folded rocks demonstrates hinge line parallel extension, although hinge line parallel contraction may be equally likely in folded rocks. The arcuate trends in parts of the central Appalachians may help determine whether hinge line parallel contraction or extension has occurred. Where the Appalchians are convex toward the craton, as in central Pennsylvania, hinge line parallel extension may be more likely. Where the Appalachians are concave cratonward, as in southern West Virginia and western Virginia, hinge line parallel contraction is more likely.

Using our values of shortening, and reasonable estimates of λ_2, the angle (θ) between the vertical X strain axis and the surface of no finite longitudinal strain is determined graphically using a Mohr diagram for three-dimensional strain (Ramsay, 1967, p. 152). The points along the line $\lambda' = 1$ on the Mohr diagram yield values of points on the surface of no finite longitudinal strain, in degrees from the X and Y strain axes. These values can be plotted in equal area projection to determine the shape of the surface of no finite longitudinal strain: conical with hinge line-parallel strain, planar without (Ramsay, 1967, Fig. 4-21). The minimum bed dip at which extension fractures will open is measured along the east-west axis of the projection.

RESULTS

Table 1 lists the predicted bed dips at which extension fractures will begin to open for given values of shortening and $\lambda_2 = 1$. These values were calculated using equation (5) and are complements of the angles (θ) between the x strain axis and the surface of no finite longitudinal strain. Some of the predicted dips were checked by Mohr diagrams for three-dimensional strain. With no change in length parallel to the hinge line, the minimum bed dip at which extension fractures will begin to open is 45 degrees. That finding is consistent with the results of finite-element modeling of viscous layers (Dieterich and Carter, 1969), and other work summarized by Perry (1978, p. 524). Where there is $\pm.1(e_3)$ hinge line parallel strain, the minimum bed dip at which extension fractures will open does not differ significantly from the values in Table 1.

Table 2 lists measurements of shortening for anticlines in West Virginia, determined from cross-sections by the sinuous bed method or from estimates in the literature. With the exception of the Burning Springs anticline, in Wood and Wirt Counties, West Virginia, no anticlines west of Gwinn's (1964) high plateau, which have structural relief less than 300 feet (90 meters), should show extension fracturing unless faulting in the subsurface has created abnormally high limb dips. Fault imbrication in the subsurface of the Burning Springs anticline has caused dips as steep as 68 degrees (Shockey, 1954). The Arches Fork and Wolf Summit anticlines may also show extension fracturing, since Gwinn (1964) indicates that they have between 300 and 800 feet (90 and 240 meters) of relief, intermediate between anticlines of the high plateau and those of western West Virginia.

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PERCENT SHORTENING	MINIMUM LIMB DIP
1%	45 degrees
2%	45 degrees
3%	46 degrees
4%	46 degrees
5%	46 degrees
10%	48 degrees
1.5%	50 degrees
20%	51 degrees
25%	53 degrees
30%	55 degrees
35%	57 degrees
40%	59 degrees
45%	61 degrees
50%	63 degrees

Table 1. Estimate of limb dip at which extension structures will begin to form for given percent shortening.

Anticline	Location	Shortening	Maximum Limb Dip	Source
RELIEF GREATER THAN 800 FEET (240 METERS)				
Burning Spgs.	Wood/Wirt Co.	2%	68 degrees	Shockey (1954)
Briery Mtn.	Preston Co.	6%	45 degrees	Cardea (195[illegible])
Chestnut Rdg.	Monongalia Co.	3%	12 Degrees	Mitchell (1960)
Wills Mtn.	Pendleton Co.	20%	overturned	Perry (1971)
Browns Mtn.	Pocahontas Co.	16%	overturned	Kulander and Dean (1972)
RELIEF BETWEEN 300 and 800 FEET (90 and 240 METERS)				
Wolf Summit	Lewis Co.	< 1%	5 degrees	Milner (1968)
Hiram	Harrison Co.	< 1%	< 1 degree	Cardwell (1973)
Arches Fork	Doddridge Co.	< 1%	< 1 degree	Cardwell (1973)
Warfield	Logan Co.	< 1%	< 1 degree	Cardwell (1973)
RELIEF LESS THAN 300 FEET (90 METERS)				
Mann Mtn.	Fayette Co.	< 1%	< 1 degree	Perry and Wilson (1977)

Table 2. Values of percent shortening and limb dip for selected anticlines in West Virginia.

CROSS-STRIKE STRUCTURAL DISCONTINUITIES: POSSIBLE EXPLORATION TOOL FOR NATURAL GAS IN DETACHED APPALACHIAN FORELAND[1]

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ABSTRACT

Cross-strike structural discontinuities (CSD's) are broad zones of structural disruption in the detached Appalachian foreland. At Parsons and Petersburg, eastern West Virginia, CSD's are eight to ten kilometers wide. Folds of various scales, longitudinal faults, and unmodelled gravity anomalies terminate or change style across or within the two CSD's. Both are visible on LANDSAT images. Field evidence points to their containing larger or more abundant joints than elsewhere; small normal faults, likewise. Mapping reveals little displacement across transverse faults. I favor a thin-skinned origin for the Parsons and Petersburg structures, at least in the Plateau and western Valley and Ridge provinces. The CSD's appear to have divided the allochthon into quasi-independent structural blocks. Median values of sizes and spacings of these and nine other Appalachian CSD's suggest that each contains about 1000 cubic kilometers of unusually fractured rock, and lhat they cons-tiiute 14 percent of the foreland allochthon. Cratonward extensions of CSD's can be loci of exploration in gas-producing, fractured Devonian shales of the Allegheny Synclinorium. Lattman (1958) and others suggest that short air-photo lineaments are surface expressions of fractures or fracture zones. Wells should be sited at intersections of short air-photo lineaments in the CSD's, where Devonian shales are involved in detached deformation.

PURPOSE

This paper suggests an exploration strategy for natural gas in part of the Central Appalachian Plateau province. The target is fracture porosity and fracture permeability in Middle and Upper Devonian clastic rocks (mostly shales), parts of which have been profitable or marginally profitable producers since the early 1900's (Shumaker and Overbey, 1976; Wheeler, Shumaker, and others, 1976; Schott and others, 1977; Dean and Overbey, 1978). My purpose is to suggest a method by which exploration for those structural traps can be made more efficient and effective in unexplored or presently unproductive areas where the reservoir rocks are involved in detached deformation, that is, outside the area of present commercial production in eastern Kentucky and adjacent states (Wallace and de Witt, 1975). The suggested method predicts subsurface locations of unusually intensely jointed rock volumes of drillable sizes in the central and western Plateau province, by combining (1) observations and hypotheses about major, cross-strike structural discontinuities, (2) observations and theories about short air-photo lineaments, and (3) understanding of detachment tectonics in the Central Appalachians.

CROSS-STRIKE STRUCTURAL DISCONTINUITIES

Detached forelands in several orogens contain large, cross-strike structural discontinuities (CSD's): structural lineaments or structural alignments, at high angles to regional strikes, most recognizable as alignments of disruptions in strike-parallel structural or geomorphic patterns. Wheeler and others (ms in review)[3] summarize sizes and characteristics of fifteen CSD's or groups of CSD's in the Appalachians, Ireland, and southern Chile, and cite a few published references to probably similar structures in the Canadian Rocky Mountains.

CROSS-STRIKE STRUCTURAL DISCONTINUITIES: POSSIBLE EXPLORATION TOOL FOR NATURAL GAS IN DETACHED 2 APPALACHIAN FORELAND

Eleven Appalachian CSD's have been mapped and studied in Pennsylvania, West Virginia, and Alabama (Fig. 1; see Wheeler and others (ms in review)[3] for citations of available work). The CSD's are larger and more complex than simple tear faults, fault zones, joint zones, or outcropping unconformities rotated to steep dips. Using the data summarized for three orogens by Wheeler and others (ms in review)[3], Wheeler (ms in preparation) estimates median CSD's to be about 3.5 kilometers wide, at least four kilometers deep, and at least 70 kilometers long, with a centerline spacing of about 25 kilometers (Fig. 2). Such CSD's would each contain at least 980 cubic kilometers of rock, and would include 14 percent of that part of the allochthon that they traverse. Thus CSD's can be major and fundamental parts of the architecture of a detached foreland, as volumetrically important as most longitudinal folds.

In northeast West Virginia, the Parsons and Petersburg structural lineaments are two CSD's that contain structurally disrupted rock (Fig. 3; Wheeler and others, 1974; Wheeler and Sites, 1977). The Parsons lineament was detected by Gwinn (1964), Woodward (1968), and Rodgers (1970, Plate 1A). Only Rodgers estimated its orientation and size correctly. Mapping and other work by Henderson (1973), Mullennex (1975), Trumbo (1976), Moore (1976), and Wheeler (unpublished results) defined the lineament as trending about N50W through Parsons, and as at least 55 kilometers long, ten kilometers wide, and probably three kilometers deep (Wheeler and Sites, 1977). Part of the lineament appears clearly on LANDSAT imagery. Major folds and longitudinal reverse faults, intermediate-scale folds, and outcrop-scale folds associated with northwest-vergent shear zones terminate or change abundance, size, shape, or orientation across or within the Parsons lineament (Wheeler, 1975; Wheeler, Trumbo, and others, 1976). Bends and ends of axial traces on the Geologic Map of West Virginia (Cardwell and others, 1968) are weakly consistent with possible extension of the Parsons lineament northwest about to the Ohio River. Mapping by T. Wilson (dissertation in progress)[4] and Rader and Perry (1976) is more strongly consistent with the lineament's extension southeast through the Valley and Ridge province. Thus the total possible length of the Parsons lineament is on the order of 250 kilometers.

Woodward (1968) noted the Petersburg lineament. Mapping and other work by Wilson and Wheeler (1974), McColloch (1976), Sites and others (1976), and R. Sites (dissertation in progress)[5] defined the lineament as trending about N85E through Petersburg, and as at least 80 kilometers long, 8 kilometers wide, and probably 5 kilometers deep (Wheeler and Sites, 1977). The Petersburg lineament has the same effects on folds and faults of various scales as does the Parsons lineament. It too appears clearly on LANDSAT imagery. Its total possible length is speculative. Mapping and structural analysis by J. LaCaze (Master's thesis in progress)[6] may determine whether the Petersburg lineament crosses the Appalachian structural front into the Plateau province, to intersect the Parsons lineament as suggested by Kulander and Dean (1978a).

I know of no evidence requiring either lineament to extend deeper than the basal detachment (Lower Cambrian Waynesboro shale) in the Plateau or western Valley and Ridge provinces. Unmodelled, strike-parallel gravity anomalies terminate at both structural lineaments (Kulander and Dean, 1976, 1978a). Kulander and Dean, and Wheeler and Sites (1977) conclude that the anomalies are caused by structural duplication of high-density Lower Devonian and Silurian, or Cambrian and Ordovician limestones and dolomites, and that the terminations of the anomalies occur over along-strike ends of such structural duplications. Those ends occur where underlying detachments step up along strike to higher structural and stratigraphic levels, in the manner suggested by Gwinn (1964) and figured by Harris (1970). Kulander and Dean (1978a) mapped unmodelled magnetic anomalies, and attributed them to susceptibility differences in the crystalline autochthon. Their total-intensity magnetic map does not appear to me or to Kulander (oral communication, 1978) to show that those susceptibility differences are caused by cross-strike faults in the autochthon, at least in the Plateau and western Valley and Ridge provinces of northern and central West Virginia. Balanced cross sections (Perry, 1971, 1975; R. Sites, dissertation in progress)[5] northeast and southwest of the Petersburg lineament in the western Valley and Ridge province show no evidence for abrupt change in basement depth along strike.

However to the southeast the magnetic map of Kulander and Dean shows magnetic relief of about 300 gammas across the southeastward extension of the Parsons lineament into Rockingham County, Virginia, in the eastern Valley and Ridge province. Pilant and Robison (1977) mapped magnetic lineaments in the Piedmont province of Virginia. Their magnetic lineaments parallel the Parsons lineament, and the northeasternmost is on trend with the Parsons lineament (W. Pilant, oral and written communications, 1977). Further, in southeast West Virginia, Pennsylvania, Alabama, Chile, and Ireland, CSD's resembling the Parsons and Petersburg lineaments either extend into exposed crystalline rocks, include volcanic centers or metal deposits, or pass through saddles or abrupt terminations of magnetic anomalies (see work referenced by Wheeler and others, ms in review;[3]

R. Shumaker, maps in preparation). Finally, Wheeler (ms in preparation) examined the orientations and minimum lengths of CSD's as a class, and concluded that those properties were probably more influenced by cratonic structures activated or reactivated under advancing detached blocks, than by structural processes originating within the allochthon.

Therefore I assume conservatively that the Parsons and Petersburg structural lineaments do not extend beneath the deepest detachment in the Plateau and western Valley and Ridge provinces, but may-well do so further southeast or east.

Neither the Parsons nor the Petersburg lineament exhibits significant transverse faults: they are not fault zones. However, they probably do contain unusually fractured rock (Wheeler and others, 1974; Wheeler, Trumbo, and others, 1976). Steffy (1976) found that water-well yields in part of the Parsons lineament were marginally lower than those outside the lineament, which is inconsistent with increased fracturing. However Holland and Wheeler (1977) found that systematic joints are larger in one road cut inside the lineament than in the same rocks in two similar road cuts bracketing the lineament (Fig. 4). Small mappable normal and strike-slip faults and longitudinal joints are more abundant within the Petersburg lineament than northeast or southwest of it in the same map units (McColloch, 1976; R. Williams, oral communication, 1976; R. Sites, dissertation in progress).[5]

In summary, I suggest that CSD's like the Parsons and Petersburg structural lineaments are very large volumes of unusually fractured rock, that they extend as deep as the basal detachment under the Plateau and western Valley and Ridge provinces, and therefore that they are promising rock volumes in which to seek fractured gas reservoirs.

SHORT PHOTOLINEAMENTS

Lattman (1958) defined photogeologic fracture traces as air-photo lineaments less than one mile long, and suggested that they are traces of joints or small faults. Since then many geologists have used the working hypothesis that short photolineaments one to several kilometers long are geomorphic expressions of vertical or nearly vertical zones of unusually fractured rock: joints or joint zones, or small faults or fault zones. The hypothesis is usually tested either by direct examination of exposures on the photolineament, or by a statistical test to determine whether water wells on or near the photolineament have significantly higher yields than do wells off or far from the photolineament (for example, see Gold and Parizek, 1976). In and near the Cottageville gas field, which produces from the Devonian shales in western West Virginia (Fig. 3), D. Jones (Master's thesis in progress)[7] found higher water-well yields within 30 meters (100 feet) of lineaments on low-altitude aerial photographs. Steffy (1976) found a similar result in outcropping Devonian sandstones, siltstones, and mudstones near Parsons.

The gas-bearing Devonian shales lie in the interval from about 0.6 to 1.4 kilometers (2000 to 4500 feet) below ground level in the gas fields of western West Virginia and eastern Kentucky. Eastern equivalents of those rocks lie about 0.4 to 2.3 kilometers down (1300 to 7700 feet) in the eastern Plateau province (Cardwell, 1973; Wallace and de Witt, 1975; Patchen, 1977). In most areas, the most productive rocks lie in the lower parts of those intervals (Patchen and Larese, 1976; Martin and Nuckols, 1976; Bagnall and Ryan, 1976; Patchen, 1977). Therefore to use short photolineaments as an exploration tool for the Devonian shales and their eastern equivalents, I make the common assumption that the single or multiple joints, high-angle faults, or other fractures that probably underlie the photolineaments form vertical fracture zones. I further assume that the depths of the fracture zones approximate the lengths.

If those two assumptions are correct, then air-photo lineaments one to several kilometers long overlie attractive fracture porosity and fracture permeability in and above the Devonian shales. Still more attractive are intersections of two or more such photolineaments.

EFFECT OF DETACHMENT

I have assumed conservatively that CSD's like the Parsons and Petersburg structural lineaments are confined to the allochthon, at least in the Plateau and western Valley and Ridge provinces. Also, it seems likely that the orientations, locations, or both, of high-angle fracture zones are different below than above a detachment fault: fracture zones formed before the detachment would be beheaded, with tops no longer above or connected to bottoms. Fracture zones formed during or after detachment may change orientation, size, or position across the detachment, to reflect differing magnitudes and orientations of stresses in the two regimes. For example, E. Werner (oral communication, 1978)[8] found differences in orientations and other characteristics of systematic joints across the probable outcrop of a southeast-dipping splay fault on the west limb of the

Burning Springs anticline in western West Virginia. Wilson and others (ms in preparation)[9] find different orientations of slickensides and other natural fractures in a cored interval above the Pine Mountain thrust than in the thrust zone.

Therefore both CSD's and short air-photo lineaments are most likely to be useful guides to subsurface fracture zones in an allochthon, if the deepest detachment is in or below the interval of interest, and if there are no major shallower detachments. In Fig. 5, to locate larger or more abundant fractures in pay zone 1 than those produced by detachment itself, one should drill east of the point T1. To locate more fractured rock in the shallower pay zone 2, one may drill as far west as T2.

EXPLORATION STRATEGY

The three structural criteria described above can be combined to suggest an exploration strategy for gas in structural (fracture) traps in the Devonian shales of the' allochthonous part of the Plateau province of the Allegheny Synclinorium (Fig. 6).

(1) Wells should be drilled east of the trace of point T2 (Fig. 5) across the ground surface. That trace is the upward projection of the line along which the deepest underlying detachment cuts the top of the interval of interest, in this case the base of the Lower Mississippian Berea Sandstone (Patchen and Larese, 1976; Bagnall and Ryan, 1976; Patchen, 1977). In some areas, most production may be expected from an eastward-thickening interval in the lower part of the Devonian shales, such as Brown Shale Zone II of Martin and Nuckols (1976). In such areas, the western limit of recommended drilling moves east, to a line analogous to the trace of point T1 of Fig. 5.

(2) In the area defined by the first criterion, wells should be drilled within the map boundaries of a cross-strike structural discontinuity. Examples are the Parsons and Petersburg structural lineaments in northern West Virginia, the Modoc, White Sulphur Springs, and Covington lineaments in southern West Virginia, and the Tyrone-Mount Union, Everett-Bedford, and McAlevys Fort-Port Matilda lineaments in western and central Pennsylvania (see references listed by Wheeler and others, ms in review).[3] Most CSD's now known in the central Appalachians have been found and studied in the eastern Plateau and Valley and Ridge provinces. The westward extent of most CSD's into areas of most likely commercial production is feasible but presently speculative. However, some mapped CSD's correspond to some of the structural lineaments of Gwinn (1964), Rodgers (1963), and Rodgers (1970, Plate 1A), though not to Woodward's (1968) interruptions in strike. Some of Gwinn's and Rodgers' structural lineaments occur in the western plateau province, so CSD's may also extend that far west. Further, many data with which to extend CSD's to the west are now available: well logs, structure contour maps, LANDSAT imagery, and maps of residuals from terrain-corrected Bouguer anomalies (Wheeler and others, ms in review;[3] Kulander and Dean, 1978a). Wheeler, Trumbo, and others (1976) suggested that the Parsons structural lineament may be confined below the relatively massive and stiff Mississippian and Pennsylvanian rocks, with the thick, relatively soft Middle and Upper Devonian rocks flowing to insulate surface structures from reflecting underlying CSD's. However subsurface and geophysical data like those just listed should permit one to penetrate any such insulating effect.

(3) In the areas defined by the first two criteria, wells should be drilled on intersections of air-photo lineaments one to several kilometers long.

CONCLUSIONS

Wells seeking gas in fractured Devonian shales should be sited at intersections of short air-photo lineaments, in CSD's, where the shales are involved in detached deformation. That exploration strategy is applicable at least as far west as the Mann Mountain anticline in southern West Virginia and the Burning Springs anticline in central West Virginia (Perry (1978) and Perry and Wilson (1977) suggest that the Mann Mountain anticline is the western limit of detachment in the lowest Devonian shales, and Rodgers (1963; 1970, p. 19-21) summarizes arguments that the Burning Springs anticline is the western limit of detachment in Silurian rocks). Eight partly mapped CSD's in West Virginia and Pennsylvania (Fig. 1) can probably be extended west into the central and western Plateau province. However, because the most important present gas production from Devonian shales is in southwestern West Virginia and eastern Kentucky, where the shales may be autochthonous, the exploration strategy presented here may be most useful in exploration of new areas in the eastern Plateau province of southern and central West Virginia, in northern West Virginia, and in western Pennsylvania.

One or more of the three criteria developed in this paper may be used alone, or might be combined profitably with other methods of predicting locations of structural (fracture) traps. Examples of such possible traps include (1) late-tectonic bed-extending fractures in steeply-dipping beds (Berger, Perry, and Wheeler, ms in review),[10] (2) stratigraphically confined, porous fracture facies (Negus-de Wys and Shumaker, Shumaker, and Wilson and others, mss in preparation),[11] (3) westward propagation of fracturing associated with a CSD, into autochthonous rocks (Wheeler, Trumbo, and others 1976), (4) westward propagation of detachment-related fracturing by differential shortening distributed over a vertical interval, west of actual recognizable detachment (Shumaker, 1978, and Wilson and others, ms in preparation),[11] (5) fracturing in bottoms of detached sheets (Harris and Milici, 1977; Milici and Statler, 1978;[12] Wiltschko, 1978b), (6) fractures in footwalls of ramps (Wiltschko, 1978a),[13] (7) upward propagation of anomalously-oriented joints or joint zones formed in response to distorted stress trajectories over longitudinal basement faults, such as those bounding the Rome trough in West Virginia (Kulander and others, 1977; Kulander and Dean, 1978b;[14] Advani and others, 1977), (8) fracture-porous CSD's containing gas sealed beneath an overriding detachment (Gold and others, 1978;[15] W. Bagnall (oral and written communications, 1976 and 1977) suggested that slickensided faults may seal porous fractures), and (9) bed-extending fractures on anticlinal crests, especially adjacent to a contact with an overlying much softer layer (Tapp and Wickham, 1978).

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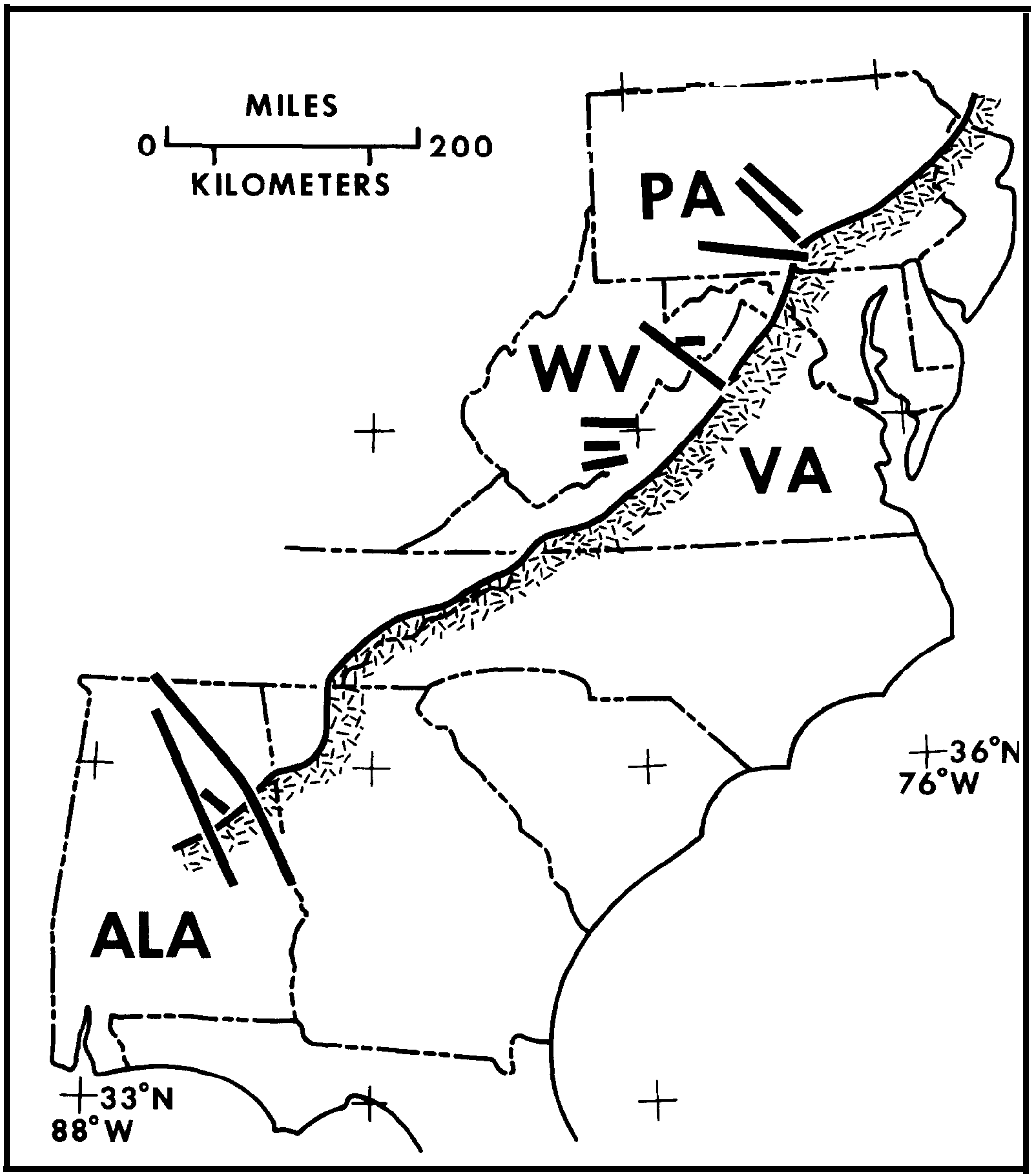

Fig. . Eleven Appalachian cross-strike structural discontinuities. Heavy solid lines: structural discontinuities. Medium-weight solid line and pattern: northwest limit of exposed crystalline rocks. Approximate locations of discontinuities from Drahovzal (1976), Gold and Parizek (1976), Dean, Kulander, and Williams (ms in review)[1,7] and Wheeler and Sites (1977). Note that only four areas have been explored in detail for CSD's, to my knowledge.

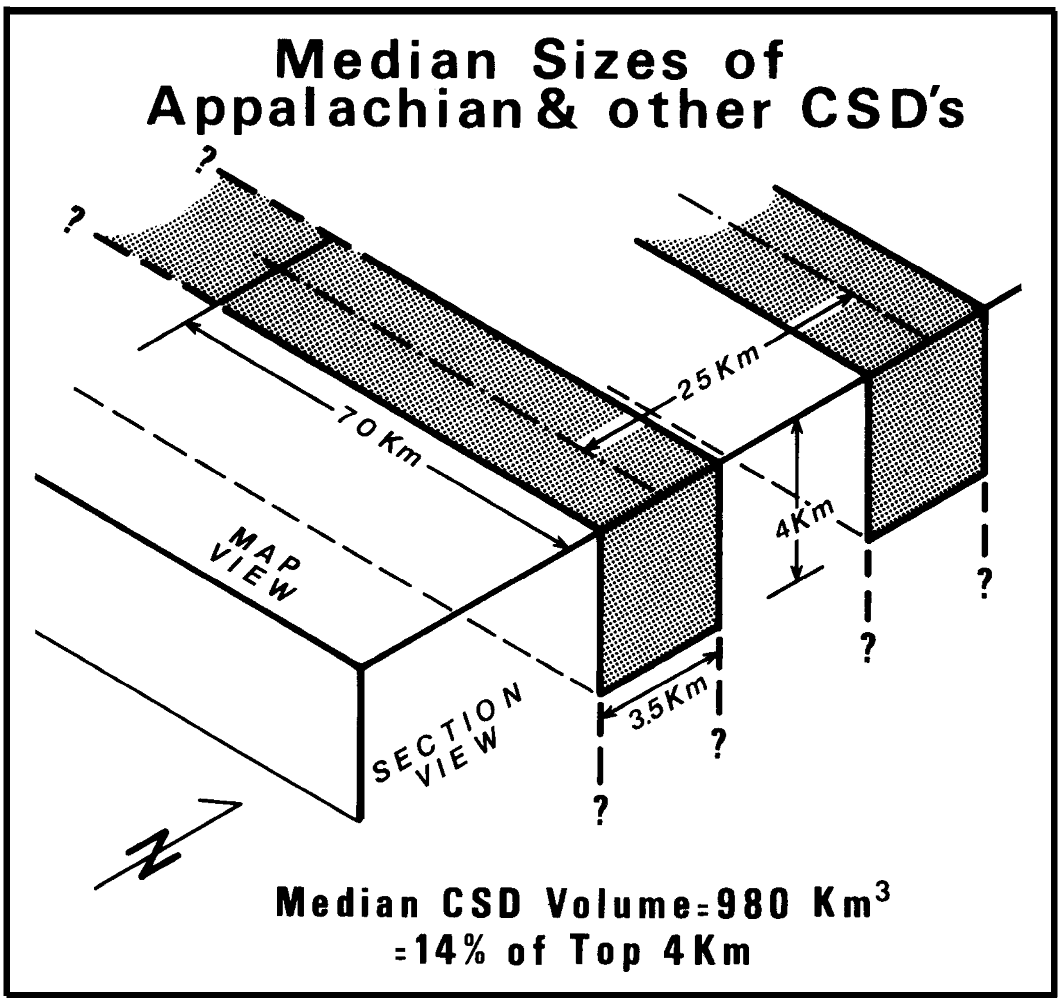
Median Sizes of
Appalachian & other CSD's
70 Km
25 Km
4 Km
3.5 Km
MAP VIEW
SECTION VIEW
N
Median CSD Volume = 980 Km^3
= 14% of Top 4 Km

Figure 2.

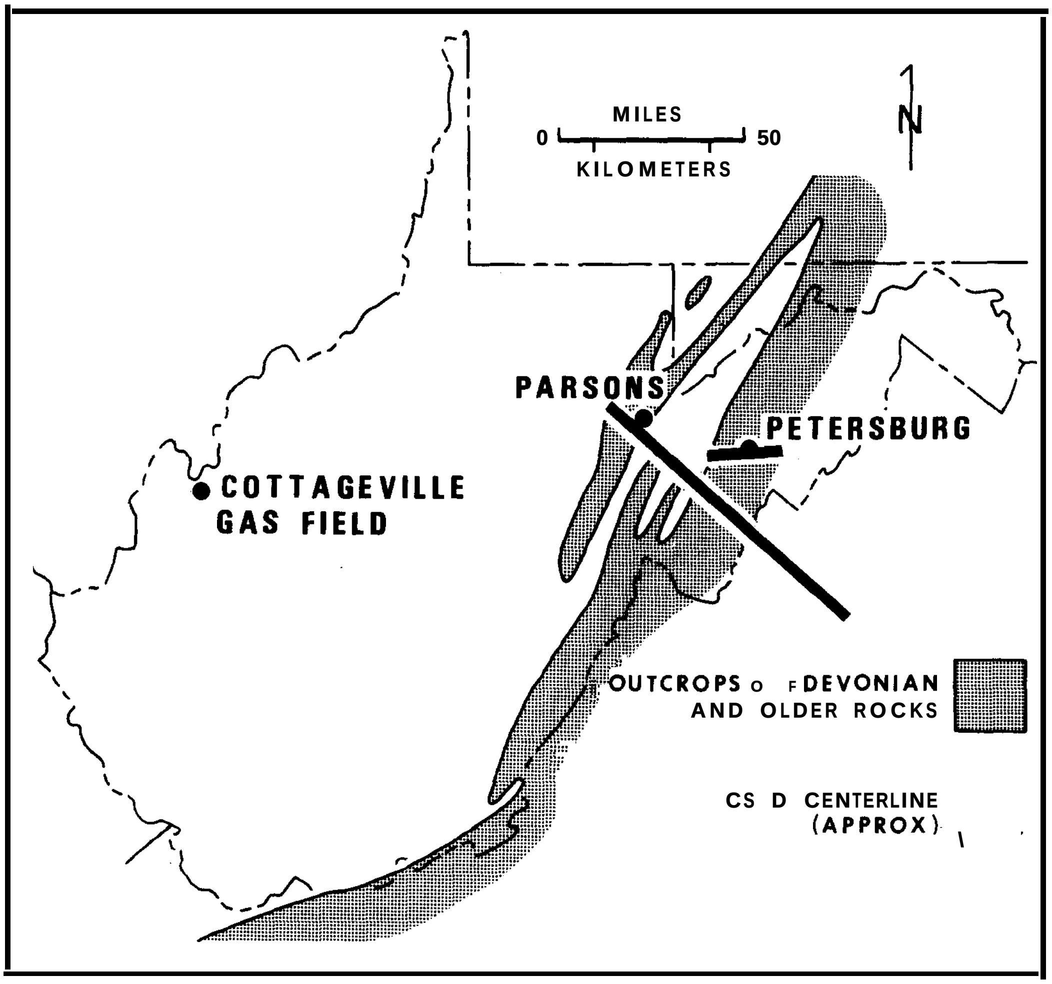

Fig. 3. Index map of West Virginia,

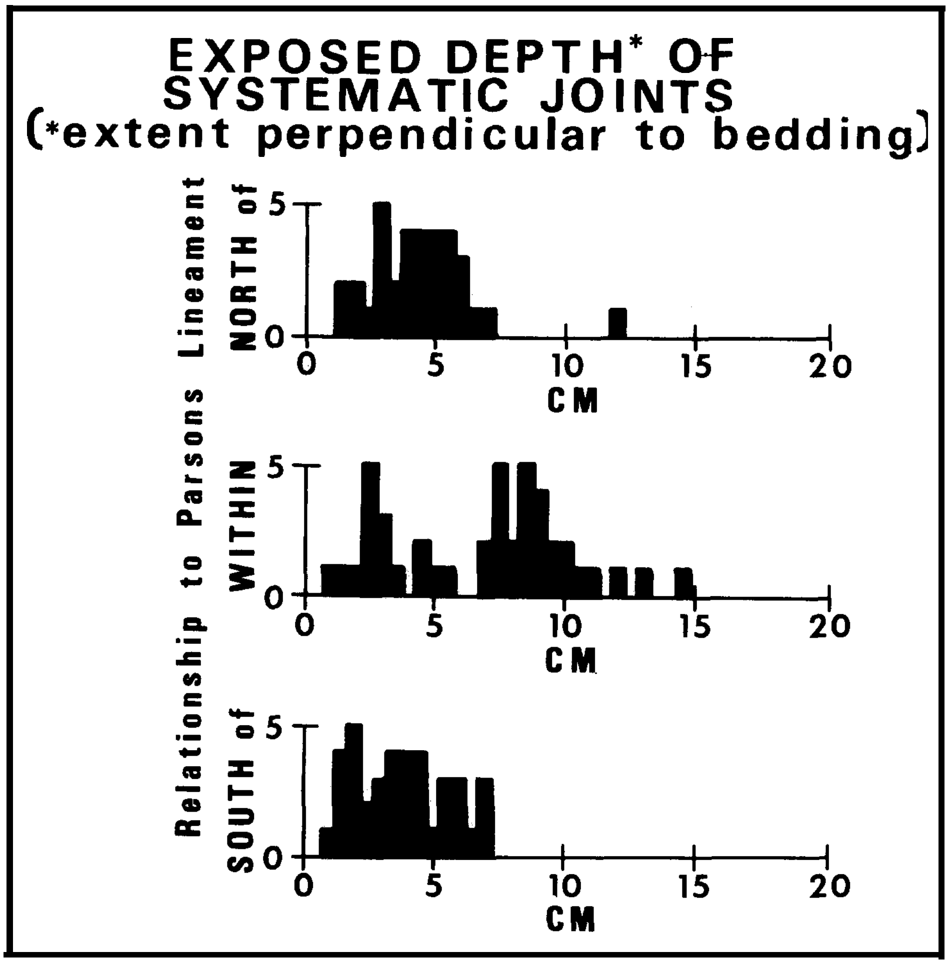

Fig. 4. Histograms of exposed depths of systematic joints in three road cuts in and bracketing the Parsons structural lineament. Joints within lineament differ from those north and south of it at significance level less than 0.001. Joints north and south of lineament do not differ significantly (Holland, 1976).

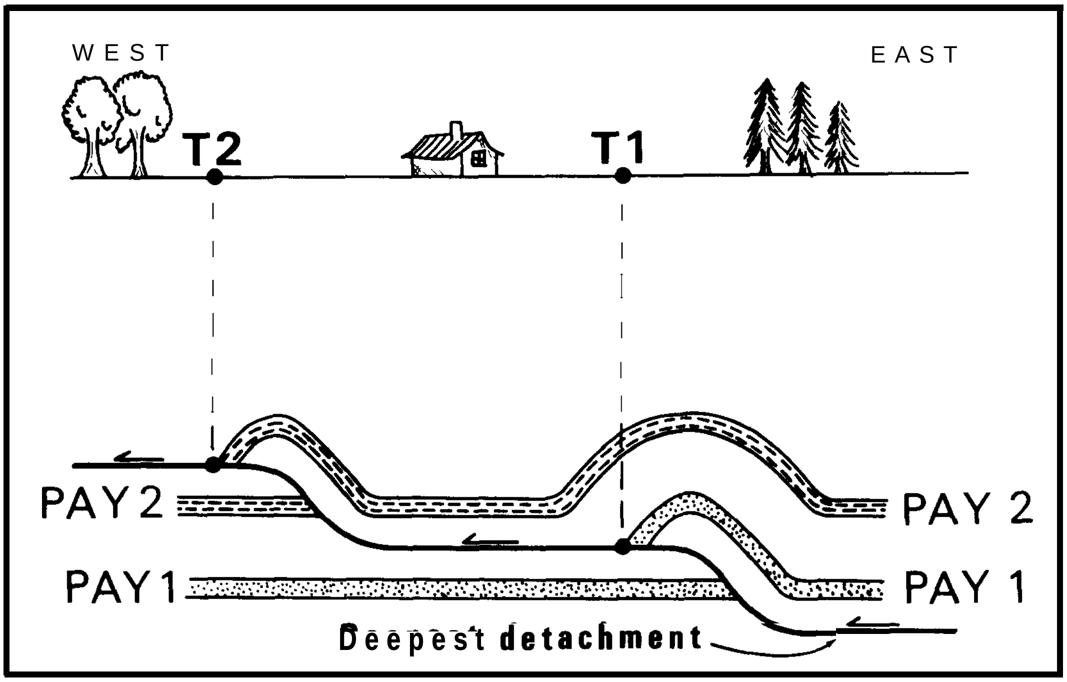

Fig. 5. Vertical cross section schematizing the westward limits of advisable drilling to locate detachment - related fracture zones (see text).

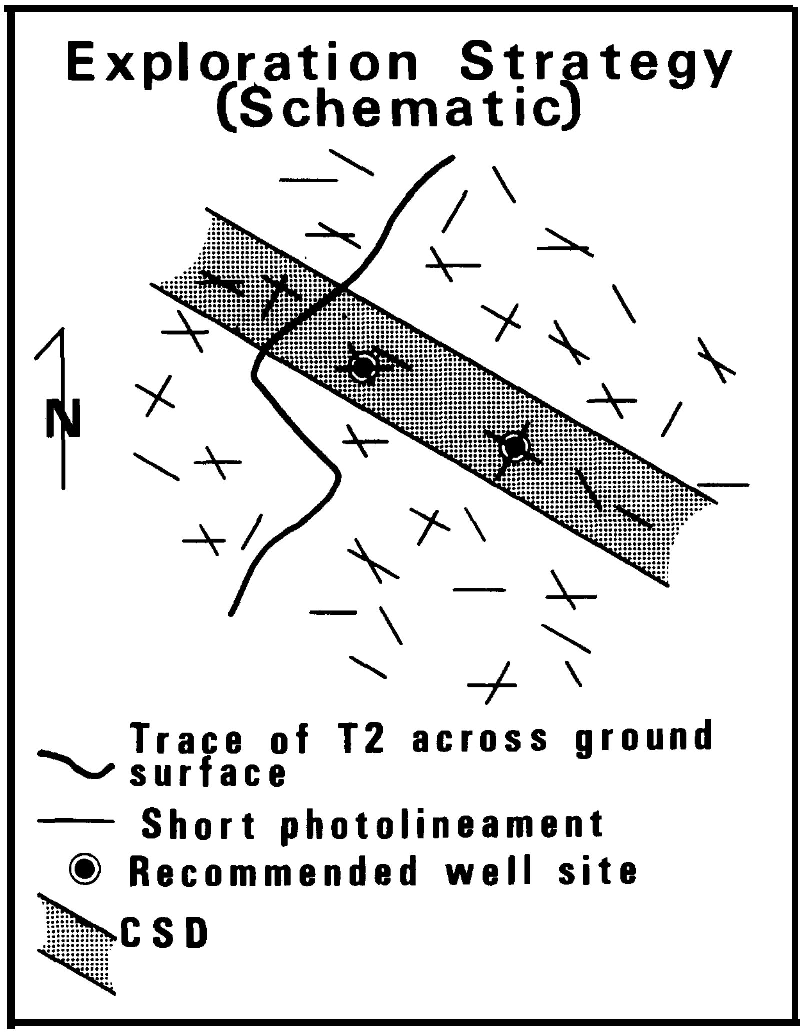
Exploration Strategy
(Schematic)
N
Trace of T2 across ground surface
Short photolineament
Recommended well site
CSD

Figure 6.

FRACTURE PATTERNS ACROSS THE BURNING SPRINGS ANTICLINE IN WEST VIRGINIA PRELIMINARY INVESTIGATION

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ABSTRACT

Fracture patterns derived from outcrop measurements and photolineament patterns derived from mapping a Landsat image in the area of the Burning Springs anticline were studied. Results indicate some differences between patterns to the east and patterns to the west of the anticlinal axis implying some differences in the stresses active on opposite sides of the anticline.

INTRODUCTION

The Burning Springs anticline (Fig. 1 and 2) is generally considered to be the western limit of major detached deformation in northern West Virginia (Rodgers, 1963; Gwinn, 1964). The characteristics of fractures were investigated to determine changes across the axis of the anticline. In this study, two avenues of investigation are being pursued: 1) mapping of photolineaments, and 2) measurements of fractures in outcrops along traverses across the anticlinal axis. This work is in progress; one traverse of 11 outcrops along U. S. Highway 50 has been completed and one photolineament map from winter Landsat imagery has been produced (Fig. 2). These preliminary data were analyzed and the results are reported herein.

PHOTOLINEAMENTS

Photolineaments have been considered as being indicative of regional fracture patterns. A considerable body of literature compares various types of photolineaments with outcrop fracture orientations. For example, Hodgson (1961) in the Colorado Plateau, Babcock (1974) in Alberta, Canada, and Hough (1960) on Chestnut Ridge in West Virginia all found a degree of agreement between orientation patterns of the outcrop fractures and of photolineaments.

To obtain an overview of photolineament patterns in the area, a map was prepared from a Landsat band 7 (0.7-1.1 μ) image enlarged to a scale of 1:250,000. Procedures used were consistent with the preparation of a type 2 map (see Werner [1977] for details); that is, the longer, more prominent photolineaments were chosen.

Fig. 2 shows the approximately 900 km of photolineaments overlain on a map of fold axes derived from Cardwell and others (1968). A statistical analysis of the data shows little, if any, significant difference between the autochthonous and allochthonous areas, respectively west and east of the axis of the Burning Springs anticline. Although the distribution of orientation patterns (Fig. 3) shows a slightly greater preferred orientation of the photolineaments in the western part of the map, the difference is not significant when tested by either the Wald-Wolfowitz runs test (see Conover, 1971, p. 350) or the Kuiper (1960) modification of the Kolmogorov-Smirnov test. Lengths do not differ materially between the two areas (west: mean length = 16.0 km, standard deviation = 7.8 km; east: mean length = 15.8 km, standard deviation = 7.9 km). Although total lineament density (length per unit area) appears to be much less for the western area than expected (23% of the total lineament length in 40% of the total area), such differences may be an artifact created by image variations. The apparent difference of the lineament density, however, should be further investigated because of implications in terms of total fracture porosity for the area. If the differences in photolineament density are real, then this could indicate different fracture densities in the two areas.

OUTCROP FRACTURES

Observations were made at 11 outcrops spaced -1.5 km apart. Fig. 4 shows the locations of the outcrops relative to the structural cross-section of the anticline. Data consist of 11 variables measured where possible for each of 247 fractures: strike, dip, lithology, bed thickness, whether the fracture entirely crosses a bed vertically, whether a fracture crosses the bedding surface either above or below the bed in which it was measured, whether the fracture is open or closed, whether the fracture is single or multiple, spacing between fractures of the same approximate orientation, character of the fracture surface, and horizontal curvature of the fracture. All accessible natural fractures were measured at each outcrop,regardless of whether they were systematic or nonsystematic joints or faults. No fractures were recognizable as faults in this area, but the relatively poor exposures made it impossible to be certain of the fracture type in all cases. The data were then divided into two groups on the basis of lithology: 1) sandstone and siltstone, and 2) shales and coals. Statistical tests were run on these groups to determine if there is any significant difference between data collected to the east and to the west of the anticline for several of the variables. These results are shown in Table 1. A comment is required here regarding the validity of results of statistical tests, particularly when many are done on the same data set. The significance level of a statistical test indicates the likelihood that the conclusions drawn might be in-error. Thus with a significance level of 0.10, there is a likelihood of 10% that the stated result is incorrect. Because a significance level of 0.10 was used and 32 tests were done, it is likely that 3 or 4 of the test results are spurious (i.e., 32 tests × 0.1 probability of error = 3.2 erroneous results). This situation must be kept in mind during the following discussion of results.

Orientation rosettes of fracture strikes from the individual outcrops are shown in Fig. 5. No clear preferred orientation appears in the diagrams to differentiate the eastern from the western part of the traverse. The statistical test (line 1, Table 1), however, shows that there is a difference in strike distributions across the anticlinal axis. When orientation rosettes are produced for the outcrop data subdivided both by lithology and location with respect to the anticlinal axis (Fig. 6), a comparison can be made. Data from the sand- and siltstone fractures (Figs. 6a and 6b) shows that west of the anticline fractures oriented at about N60°W are common, but east of the anticline this peak is completely absent. Conversely, a sand- and siltstone peak appears at about N60°E east of the anticline, but is not apparent to the west of the anticline. The shale and coal data (Figs. 6c and 6d), on the other hand, show a direct reversal of these relationships. Also, east of the anticline, the shale and coal fractures show a much more pronounced preferred orientation.

Statistical comparisons were also made of some of the other data. Fractures are straighter and more vertical in sandstones in the west, but fractures in shales and coal are straighter and more vertical (not statistically significant) to the east (lines 2 and 4, Table 1). Similarly, fractures in sandstones are generally smoother to the west, but shale and coal fractures appear to be slightly smoother (not statistically significant) to the east (line 3, Table 1). Of the various measures of total fracture area, some point toward greater fracturing to the east and others to the west. The two more important variables - fracture frequency and vertical extent (lines 5, 6, and 7, Table 1) - indicate a likelihood of more fracturing east of the anticline in the shales and coals. In sandstones, fractures may be more frequent (not statistically significant) but they are less likely to have as great a vertical extent in the east. In general, then, although fractures are spaced somewhat more closely to the east, they tend to be more regular to the west.

COMPARISON OF PHOTOLINEAMENT AND OUTCROP FRACTURE ORIENTATIONS

There are many examples in the literature of cases in which fracture and photolineament orientation patterns are generally similar; yet, at the same time, examples of the converse condition also abound. The data collected to date for the Burning Springs anticline appear to contain both conditions. Although the orientation patterns of the photolineaments to the east and the west are not statistically different, the patterns of the outcrop fractures are. Hence, it might be expected that the fracture orientation distributions on one side or the other of the axis might be different from the respective photolineament orientation distributions. This is in fact the case. Use of the Kuiper (1960) modification of the K-S test indicates a clear difference in these distributions for the area to the east of the anticlinal axis. No such difference is indicated for the area west of the anticlinal axis. If these relationships persist in future work, this may indicate that the photolineaments and those fractures prominent in the western part of the area may have a common origin. If the photolineaments were basement related and outcrop fractures related to the allochthonous deformation creating the Burning Springs anticline, then the orientation distributions might differ. The amount of data presently collected is insufficient for conclusions about the cause of the fractures, particularly because the fracture data are derived from a single traverse across the anticline.

ACKNOWLEDGEMENTS

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TABLE 1. RESULTS OF STATISTICAL TESTS ON FRACTURE DATA

	Statistical Test[1]	all result[2]	all sig.[3]	sand & silt result[2]	sand & silt sig.[3]	shale & coal result[2]	shale & coal sig.[3]
1. Is the distribution of strike valaes different between the eastern and western areas?	Runs	indet.					
	Kuiper	yes	*				
2. Are fractures straighter to the vest?	M-W	no	•	yes		no	*
	K-S	yes		yes	*	no	*
3. Are fracture surfaces sore likely to be smooth to the vest?	Chi-square	yes	*	yes	•	no	
4. Are fractares sore vertical to the vest?	M-W	yes		yes	*	no	
	K-S	yes	•	yes	*	no	
5. Are fractares spaced more closely to the east?	M-W	yes	☒	yes		yes	*
	K-S	yes	☒	yes		yes	*
6. Are fractures more likely to cross beds to the east?	Chi-square	no		no	*	yes	*
7. Are fractures more likely to cross bedding surfaces to the east?	Chi-sqaare	yes		yes		yes	
8. Are fractures more likely to be compound to the east?	Chi-square	no		yes		no	*

Notes:

1. Statistical test: M-W - Mann-Whitney U-test, see Conover (1971), p. 224.
   K-S - Kolmogorov-Smirnov test, see Conover (1971), p. 309.
   Runs - Wald-Wolfowitz runs test, see Conover (1971), p. 350.
   Kaiper - Kuiper's (1960) variation of K-S test for circularly distributed data.
   Chi-square - chi-square contingency table test, see Conover (1971), p. 140, 150.
2. Result is in terms of the answer to the question stated, whether or not it is statistically significant. The answer is based on the medians for the two groaps where numeric valaes exist, and on counts of fractures falling into varioas classes for the other variables.
3. Significance is indicated by asterisk when the test is significant at a level of $\alpha = 0.10$; that is, the probability of the stated result being in error is 10% or less.

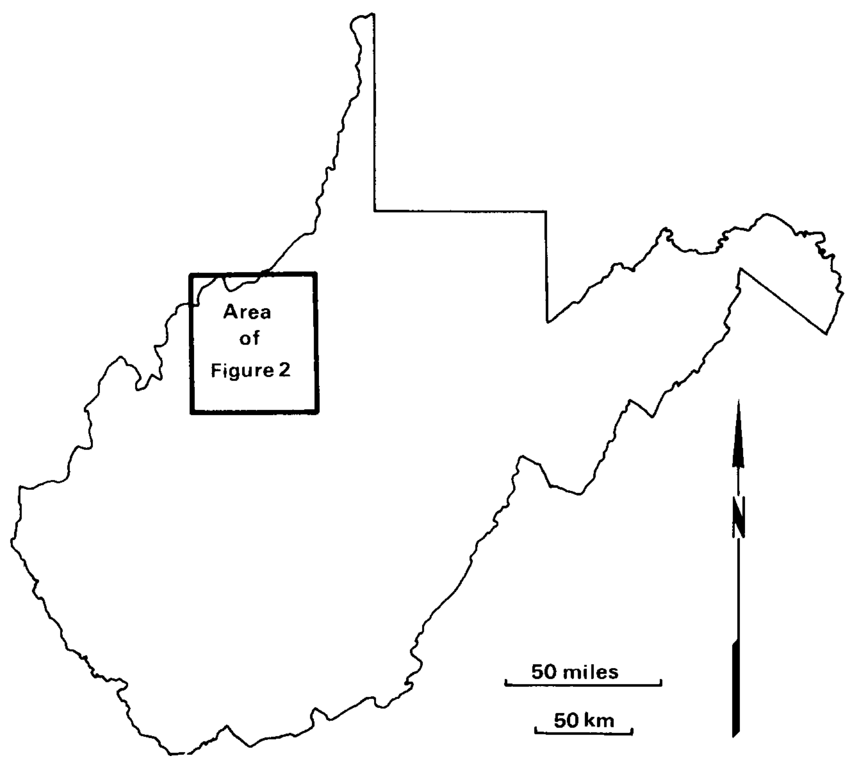

Fig. 1. Index map of West Virginia showing area of investigation.

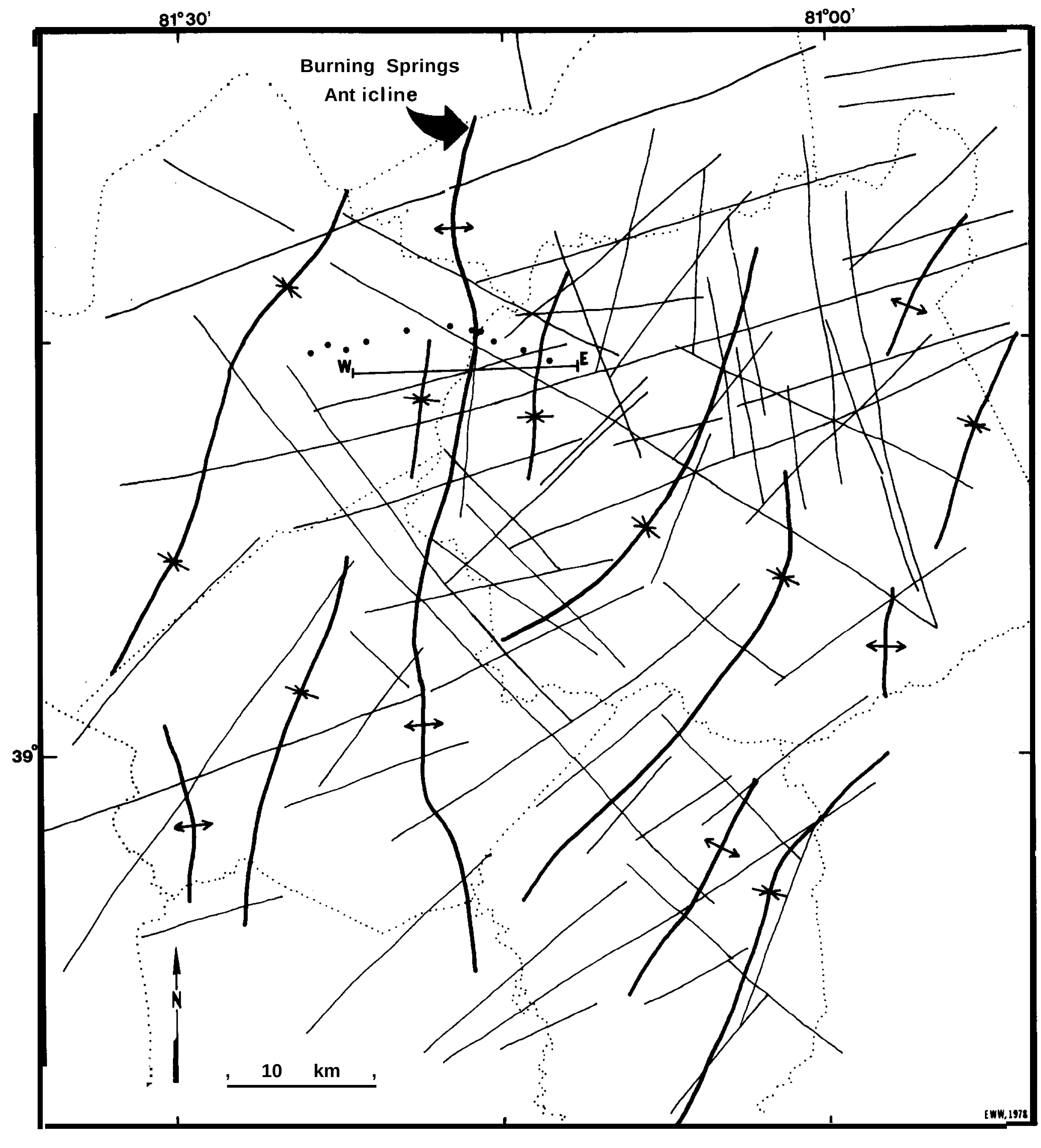

Fig. 2. Photolineaments (light lines) and fold axes (heavy lines) in the study area. Line W-E is line of cross-section shown in Fig. 4. Light dotted lines are county boundaries. Large dots indicate outcrops used for fracture measurements.

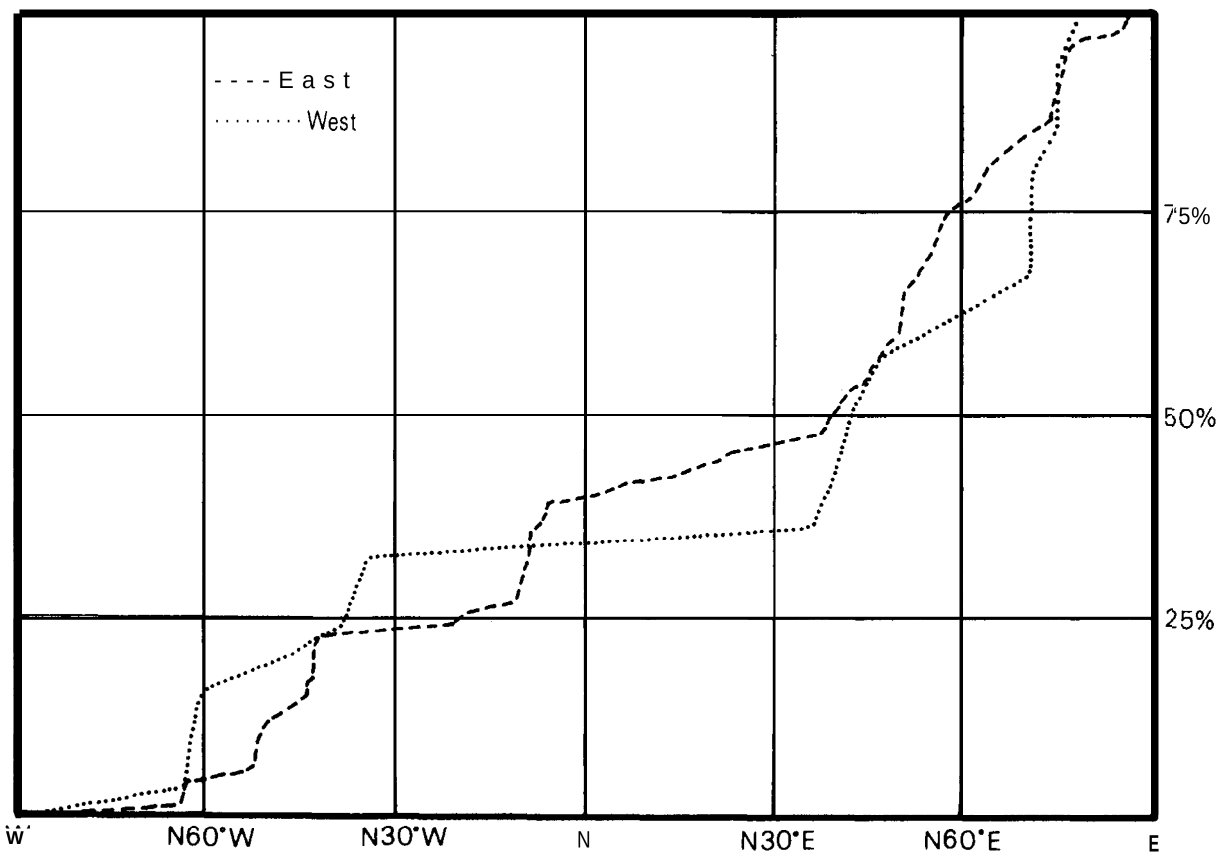

Fig. 3. Cumulative distribution by orientation of photolineament lengths shown in Fig. 2.

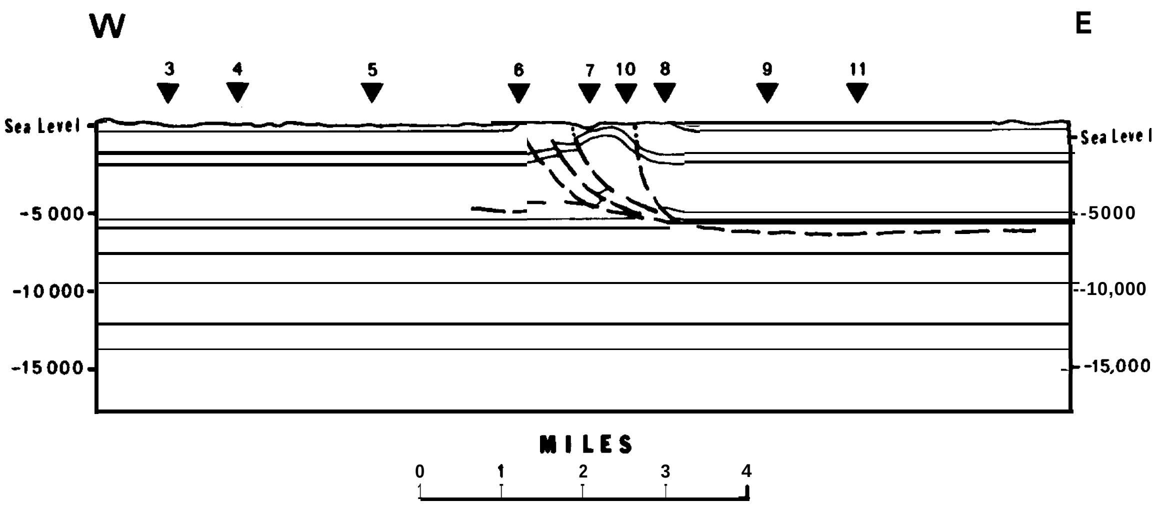

Fig. 4. Structural cross-section of the Burning Springs anticline (after Cardwell and others, 1968). Projections of 9 of the measured outcrops into the plane of the section are shown. The remaining two outcrops are to the west of the cross section.

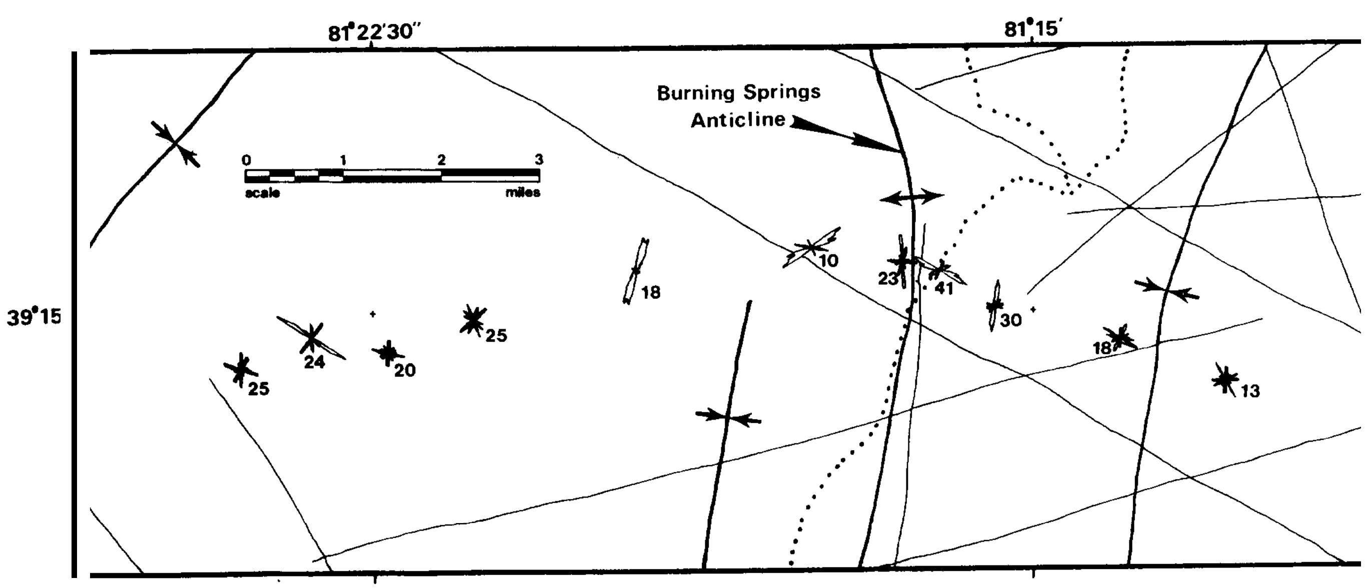

Fig. 5. Orientation rosettes of the 11 outcrops measured along U. S. Highway 50. Numbers are number of observations at each site. Photolineaments (light lines) and fold axes (heavier lines) are transferred from Fig. 2.

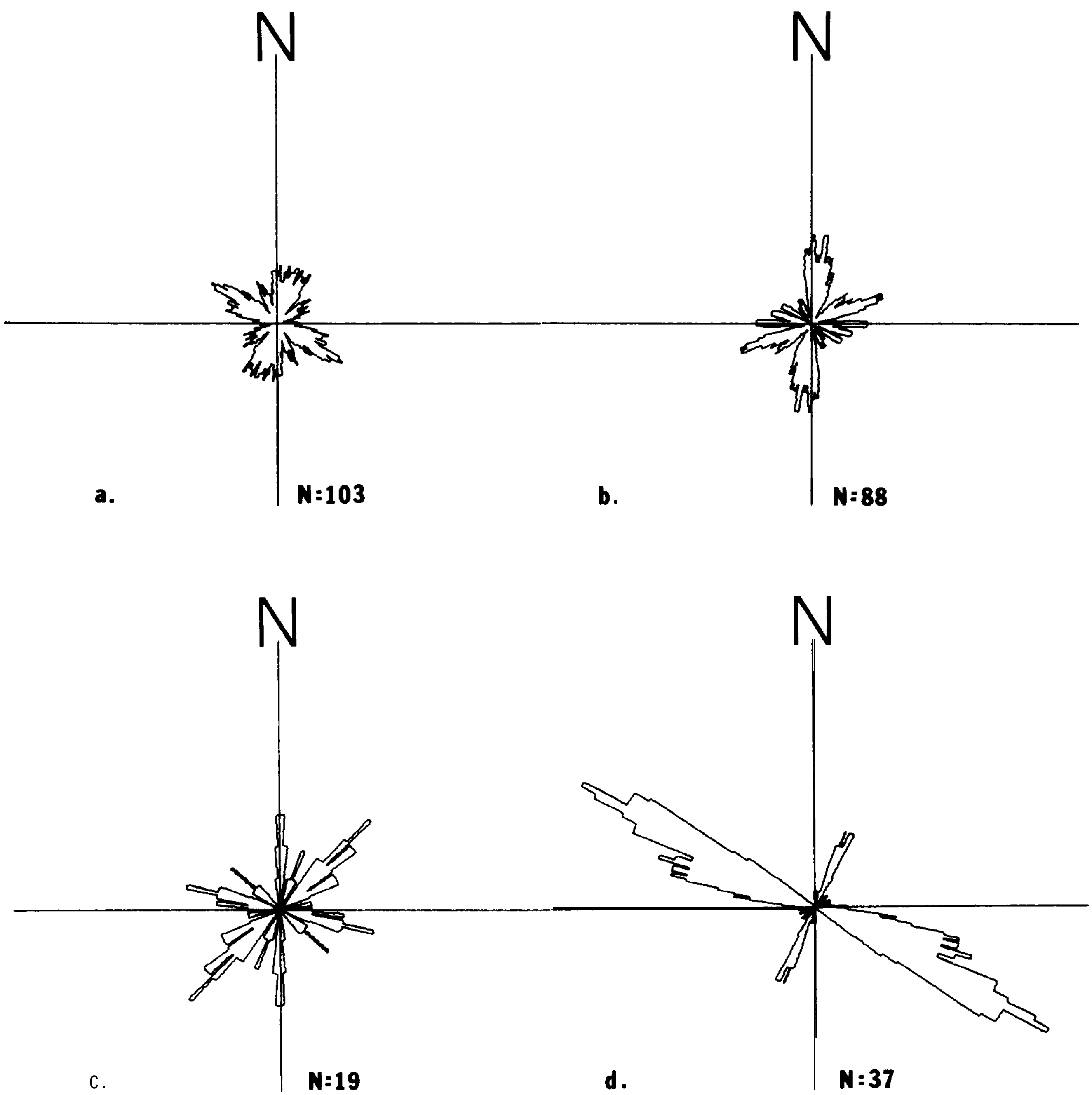

Fig. 6. Rose diagrams of fractures. a - sand- and siltstone fractures west of Burning Springs anticline; b - sand- and siltstone, on and to the east of the anticline; c - shale and coal, west of the anticline; d - shale and coal, on and to the east of the anticline.

ROME TROUGH RELATIONSHIP TO FRACTURE DOMAINS, REGIONAL STRESS HISTORY AND DÉCOLLEMENT STRUCTURES

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ABSTRACT

The Rome Trough, a major Appalachian basement structure, trends northeastward through western West Virginia from Kentucky into Pennsylvania. Its eastern flank, through Kanawha and adjoining counties is steepened by pronounced faulting. Total east margin relief from the trough axis to the crest of the central West Virginia arch is approximately 3000 meters (10,000 feet). Resulting basin configuration may have influenced past and present stress trajectories in Paleozoic rocks.

Three coal fracture domains were mapped in south-central West Virginia. Here regional systematic fractures predate folding and strike N45W and N75-85E over the Rome Trough and central West Virginia arch respectively delineating two fracture domains. To the south and southwest of Kanawha County, a west-northwest trending boundary parallels the northern flank of the southern West Virginia arch before curving to the northwest into Wayne County. This line delineates a third fracture domain. Within this domain multiple systematic fracture sets (N15W, N45W, EW) increase fracture permeability and chaotically break the coal. The boundary separating the N45W and N75-85E domains contains coal chaotically broken by fractures of both domains plus locally-developed sets. This boundary lies directly over the Rome Trough basement fault zone.

Regional stress configurations have changed across the plateau through time. Fracture domain boundaries mark prefolding regional stress boundaries. In each domain coal fractured as a brittle solid, and fractures propagated perpendicular to the greatest tension. In contrast, stresses associated with Alleghanian folding differed from those responsible for coal fracturing. Present day stresses suggested by in situ stress measurements (Overbey, 1976) indicate a still different stress configuration that changes radically over the Rome Trough. A basin-wide finite stress analysis (Advani, 1977) suggests that stress fields could be altered in rocks overlying a faulted basement. It is likely that any stress alteration related to basement structure existing during décollement folding may have affected the detachment mechanism and resulting tectonic styles.

GENERAL GEOLOGY

Basement Surface Structure

Previous studies have shown that the Precambrian surface configuration within the investigation area is dominated by three regional basement structures (Kulander and Dean, 1977, 1978). The most pronounced of these is the Rome Trough flanked to the east and south by the central and southern West Virginia arches respectively (Figure 1). Evidence to date suggests that the deepest section of the Rome Trough lies in Kanawha County, The Rome Trough's eastern flank and corresponding western limb of the central West Virginia arch are accentuated by a zone of pronounced normal faulting. Normal faults within this zone have a maximum combined throw of approximately 2135 meters (7000 feet) within a horizontal distance of several kilometers. The Rome Trough's western margin is also undoubtedly faulted. However, to date, the authors have not had access to seismic data, and this assumption remains uncorroborated. Also magnetic and gravity data do not indicate faulting of the magnitude evident on the eastern flank of the Rome Trough.

The major regional basement structures, including the basement fault zone, have pronounced magnetic and gravity signatures. Geophysical data also suggest intrabasement mafic plutons (Kulander, Dean, 1978). Moreover, preliminary gravity and magnetic calculations modeled on the faulted basement and mafic basement intrusive bodies, utilizing appropriate density and susceptability contrasts, are of the proper magnitude and reflect the existence of the fault zone (Kulander, Dean, Williams, 1978).

Paleozoic Allochthon Geology

Exposed bedrock in the study area ranges in age from lower Pennsylvanian through the Permian Dunkard Group. The region lies entirely within the Allegheny Plateau and is traversed, with several exceptions, by the low amplitude folds characteristic of this structural province. Figure 2 depicts the area1 geology above the basement fault zone including surrounding regions within the investigation area. Two distinct fold trends are evident with predominant and subordinate axial trace directions of N40-45E and N-S respectively. A form-line map of surface coals in West Virginia compiled by Shumaker (1974), depicts surface rock configuration and compliments structural data shown on the West Virginia State Geological Map (Cardwell, et al., 1968).

The allochthonous flexures with the greatest wavelength, amplitude and extent within the investigation area are the Warfield anticline and northern and southern ends of the Mann Mountain and Burning Springs anticlines respectively. These structures are characterized in the subsurface by contrasting asymmetry and tectonic thickening at and above different decollement depths.

1. Structure contour maps by Haught (1968) and Perry and Wilson (1977) show that the Mann Mountain anticline is reflected only at the comparatively shallow middle Mississippian Greenbrier level. Older lower and middle Devonian Oriskany and Onondaga strata are not involved in Mann Mountain folding (Cardwell, 1973, 1974) indicating tectonic thickening of decollement units between the Greenbrier and Onondaga formations. Moreover at the Greenbrier level the Mann Mountain anticline is markedly asymmetrical towards the west with a closure of 90 meters (300 feet), and maximum relief approaching 150 meters (500 feet).

2. The Burning Springs anticline is developed at the Onondaga level, Structural relief of 460 meters (1500 feet) is indicated at the top of the Onesquethaw (Onondaga) in the Sand Hill area. At this level the structure has a pronounced westward asymmetry. This asymmetry is enhanced by thrust faulting along the crest of the anticline and along the western flank.

3. The Warfield anticline is the major allochthonous structure formed over the deep axial region of the Rome trough, and is situated immediately west of the basement fault zone. This anticline is different from the previously described structures in several respects. The Warfield anticline is discernible not only on Greenbrier and Onondaga maps but also on a structure map of the Silurian Williamsport sandstone as well (Cardwell, 1971). Structural relief indicated for the western and eastern flanks of the Warfield structure at the Greenbrier and Onondaga levels is 150 meters (500 feet) and 300 meters (1000 feet) respectively along the southern Kanawha County boundary. In addition the structure is asymmetrical towards the east at all levels.

Structural contours on the Williamsport and Onondaga formations show vertical northeastward plunge components of over 300 meters (1000 feet) along the Warfield anticline. However, the Vertical northeast Warfield plunge component at the Greenbrier level is 90 meters (300 feet). The 300 meter Onondaga-Williamsport plunge component corresponds with the northeastward Rome trough basement deepening in Kanawha County, whereas the 90 meter vertical Greenbrier plunge component corresponds with the plunge shown on the coal form-line map of Shumaker. Fart of the 210 meter (700 feet) Warfield anticline plunge difference between Onondaga and Greenbrier levels is attributed to the fact that Greenbrier rocks rise regionally to the northeast in contrast to Onondaga strata that plunge regionally northeastward.

A regional isopach map of all strata included in the interval between the tops of the Greenbrier and Onondaga formation is given in Figure 3. These strata consist predominantly of Devonian shales that are of primary importance to eastern gas shale production. The isopach map shows a regional northeastward thickening of middle Devonian through middle Mississippian units. Similar northeastward thickening on a regional scale is shown on a statewide Devonian shale isopach map by Patchen (1977). However, local thickening associated with the Warfield anticline, and northern and southern ends of the Mann Mountain and Burning Springs structures respectively, disrupts this regional trend. Interval thickness changes also occur within other less pronounced anticlines and synclines. Also, the isopach contours indicate localized anomalous thickness changes over much of the basement fault zone.

An isopach map of the Onesquethaw stage by Cardwell (1973) shows a regional thickening of these strata over the Rome trough, In contrast, the same map shows Onesquethaw thinning over and parallel to the southern West Virginia arch. Regional thickness variations of this nature may reflect basin subsidence synchronous with deposition, in contrast to localized tectonic thickening of strata above a decollement.

COAL FRACTURES AND FRACTURE DOMAINS

The tendency of coal seams to fracture systematically into face and butt cleat (orthogonal systematic-nonsystematic fracture sets) is widely known. Regional studies of fracturing directions in coal have been done in Ohio by Ver Steeg (1942) and Pennsylvania by Nickelsen and Hough (1967). Systematic fractures in all lithologies are currently being studied throughout West Virginia, southwest Virginia and western Maryland by the authors.

Coal Fracture Domains

Systematic and abutting nonsystematic coal fractures maintain a uniform orientation throughout large areas, thereby defining distinct fracture domains (Figure 4). Coal fractures in any domain are, with few exceptions, perpendicular to bedding, and systematic-nonsystematic fractures are generally orthogonal to each other. Where regional fractures occur on fold limbs or within fault zones, they bear no consistent relationship to these structures. Where bedding dips steeply regional coal fractures are merely rotated and remain perpendicular to bedding. However, local systematic and nonsystematic coal fractures that post-date and abut against regional sets can form perpendicular and parallel to fold hinges (Figure 5), or develop due to localized stresses about fault zones. Therefore regional systematic coal fractures are, as a rule, not related to the stresses that caused folding and faulting. In eastern West Virginia regional systematic fractures in Mississippian and Devonian sandstone shale and limestone also predate folding (Dean, Kulander, 1977; Dean, Kulander, Williams, in press).

Fracture patterns in any rock type can generally be separated into distinct regional or local fracture domains. Regional coal fractures in any given domain can possess a singular trend for over a thousand square kilometers. A given domain may also be marked by systematic fractures of several distinct trends. Fracture domain boundaries can be sharp (up to 15 kilometers in width) or gradational. Domain boundaries can separate fracture domains characterized by fracture sets of distinctly different trends. Domain boundaries can also simply delineate a region where additional fracture sets uncommon to an adjacent domain occur in conjunction with one or several fracture sets common to both domains. If simultaneous fracturing occurred in two adjacent domains, the intervening domain boundary also acted as a stress boundary.

Three fracture domains, denoting different stress to fracture orientations, are evident on the coal fracture trend map of southern West Virginia.

1. Systematic fractures over the Rome trough in Kanawka, Lincoln and Boone Counties trend N45W with few exceptions and partially define one fracture domain. Systematic fractures within this domain are well formed at any outcrop and regional fracture trends vary little about the mean trend at any given outcrop. The western and northern boundaries of this fracture domain have not been determined.

2. Another distinct fracture domain is developed in coals over the central West Virginia arch to the east of the zone of Rome trough basement faulting. Within this domain fracture trends strike predominantly N75 to 85E. Again these fractures are well developed and tightly grouped about a mean trend at most outcrops. The eastern and northern boundaries of this fracture domain are currently being determined.

3. The remaining fracture domain is situated over the northern flank of the southern West Virginia arch. This domain is unique from those previously described because it is characterized by multiple systematic fracture sets. Within this domain fracture sets trend N15W, N45W, and EW, and occur together at many outcrops. Any one fracture set may be dominant at a given outcrop, and fractures of any given set are grouped more variably about a mean trend. When two or three of these fracture sets occur together, fracture permeability is increased, and the coal is chaotically fractured.

Coal Fracture Domain Boundaries

The boundary separating domains one and two is marked and abrupt. Fracture trends common to one domain change within two to fifteen kilometers to fracture trends characteristic of the adjoining domain. Systematic fracture trends common to both domains may occur together within this domain boundary. In addition local fracture sets can also be present. This domain boundary coincides exactly with the underlying basement fault zone.

The irregular east-west trending domain boundary approximating the 38° parallel separates the first and second from the third described fracture domain. The western section of the boundary is abrupt and occurs where the single fracture set common to the northwestern domain (first described) is developed in conjunction with additional fracture sets common to the third domain. The N45W fracture set occurs across the fracture boundary and is common to both domains. The eastern sector of this domain boundary is more indistinct and occurs over a wider area. This boundary sector separates fractures to the north that trend N75-85E (second described domain) from south-lying fractures of the third described domain.

Fracture Frequency - Stress to Fracture Directions

Fracture frequencies in coal of all domains are as a rule more highly variable at a single outcrop than over an entire fracture domain. Fracture frequency in glossy vitrain layers is greater than in dull fusain and durain coal layers. Fracture frequencies in bone coal are lowest of all, and bone coal fractures can differ in trend from those in associated coal seams.

All mapped systematic and nonsystematic fractures in coal reflect stresses essential to the formation of mode I fractures (Kulander, Barton, Dean, in press). In addition the morphology of coal fracture surfaces (transient and tendential features) support the assumption that all systematic and nonsystematic coal fractures are the result of brittle failure. These fractures formed in response to a principal tension acting perpendicular to the fracture plane. Prefold systematic fracture frequency could be increased by more recent fracturing utilizing an early anisotropy (perhaps microfracturing) produced by the prefold stress. Figure 6 illustrates several transient features common to systematic and nonsystematic fracture faces that have not been subject to extensive alteration.

REGIONAL STRESS HISTORY

Four distinct regional stress events can be documented for southwestern West Virginia:

1. Overbey (1976) has derived maximum surface in situ stress trajectories across West Virginia from bore hole strain gage measurements and hydraulic fracturing results. Maximum surface stress trajectories are interpreted as being approximately east-west over much of the Allegheny Plateau. However, over the Rome trough, trajectory directions change abruptly to a northeast-southwest trend, paralleling the trough axis.

2. Present day in situ stresses are obviously different than those responsible for central and southern Appalachian folding, Regional horizontal maximum stresses initiating Alleghanian folding are commonly interpreted as possessing a past orientation perpendicular to regional structural trends. In no case is this stress direction east-west or northeast-southwest. For example, recent investigations of the orientation of the pits and columns on stylolite faces cutting Greenbrier limestone in southwest Virginia and south and central West Virginia indicate maximum paleostress directions perpendicular to fold axes. (Dean, Kulander, 1978) These stylolite teeth trend N10-30W and N60-70W in areas affected by southern and central Appalachian stresses respectively.

3\. and 4. Systematic and nonsystematic coal fractures that predate folding indicate yet other maximum paleostress orientations. As previously stated, coal fractures propagated perpendicular to the principal tension. That this principal tension direction was spatially variable is indicated by the three described fracture domains. Two assumptions can be made concerning the stress configuration. 1) Fractures were initiated by a state of triaxial extension; 2) Principal compressive stresses acted perpendicular to the resultant principal extension responsible for prefold fracture development. In the latter case most principal compressive stresses would differ in orientation from stresses attributed to the Alleghenian folding process, as well as present day stresses.

One notable exception may occur to the general rule that coal fractures are unrelated to Alleghenian folding compressive stress. A coal fracture set in the domain south of the east-west domain boundary (described domain three), is oriented N15W perpendicular to the southern Appalachian structural trend and parallel to southern Appalachian Greenbrier stylolite teeth axes further east. It is possible that multiple coal fracturing in this domain may be attributed to stresses responsible for Appalachian folding. However, the authors contend that such fractures, if formed under these stress conditions, would have developed before any significant buckling occurred because of the brittle nature of the coal. A detailed investigation of transient and tendential fracture characteristics would prove or disprove this statement.

Nonsystematic fractures formed, in all cases, after geometrically related systematic sets. Systematic-nonsystematic fracture pairs maintain an orthogonal relationship. Therefore, it follows that the principal extension directions responsible for these fractures acted at different times and perpendicular to each other.

DECOLLEMENT OBSERVATIONS

Geologists have for some time accepted the fact that Paleozoic shale sections have served as detachment horizons beneath the Allegheny plateau. For example, in the area under investigation tectonic thickening and thinning of Devonian decollement strata occurs within anticlines and synclines. Also the Devonian shale regional thickening rate varies over the Rome trough basement fault zone. Some tectonic decollement thickness variations may have been enhanced by local changes in the prevailing stress field not attributable to alterations of dimensional or physical properties of any given slip horizon. For example an altered stress field could also affect the sense of movement and/or local deformational style in any decollement zone.

Discussion

The Mann Mountain anticline is asymmetrical towards the west at the comparatively shallow Greenbrier depths, and folding does not extend into Onondaga and older rocks. Perry (personal communications)[2] concludes that the Mann Mountain anticline may be the western limit of detachment at the Millboro level in south-central West Virginia. In contrast the Warfield anticline, trending parallel to, and immediately west of, the basement fault zone, is asymmetrical towards the east. In addition, the Warfield structure is reflected in rocks below the Silurian Williamsport sandstone. This indicated thickening of pre-Williamsport rocks may be explained by additional decollement activity in underlying Silurian or upper Ordovician shales. If thickening below the Williamsport level is attributed to westward decollement movement within these lower levels, one must assume that this decollement activity transported the Mann Mountain structure westward with no concomitant tectonic decollement thickening. Furthermore the eastward asymmetry of the Warfield anticline would therefore have to be attributed to underthrusting or some such mechanism. The regional isopach map (Figure 3) reflects the contrasting asymmetry of the Mann Mountain and Warfield anticlines and shows Devonian lower Mississippian thickening within these structures to be asymmetrical to the west and east respectively.

Warfield and Mann Mountain structural geometry and folded Silurian rocks in the Warfield anticline suggest eastward decollement movement into the Rome trough at the upper Ordovian-lower Silurian as well as the Devonian level. Eastward decollement movement would be facilitated by an eastward regional basement gradient that is steepened along the west flank of the Rome trough. Furthermore it follows that the limit of major westward decollement transport would lie between the Warfield and Mann Mountain structures at one or all decollement levels. The chronological development of the Warfield Mann Mountain structures is a critical untested factor in any such placement of the westward detachment boundary.

The basement fault zone, accentuating the western flank of the central West Virginia arch, is also situated between the Mann Mountain and Warfield structures. This spatial relationship suggests that the steepened gradient along the western margin of the Rome trough and Rome trough faults may have played a role in altering principal stress configurations during times past as well as the present. Any basement fault-related regional stress alteration occurring with decollement activity may have affected the decollement mechanism itself. Evidence for regional stress alterations related to the zone of basement faulting is four fold.

1. Principal stresses initiating systematic and nonsystematic coal fractures differed across the domain boundary over the basement fault zone. Here, it is probable that coal fractures in both domains formed simultaneously.

2. Present day in situ principal stresses change abruptly over the Rome trough.

3. Advani, et al. (1977) have shown by finite stress analysis that simulated stress magnitudes and trajectories are altered about a buttress such as the Rome trough basement faults.

4. The basement fault zone lies between the Warfield and Mann Mountain anticlines as does the proposed limit of western decollement movement.

Decollement Porosity

Devonian gas bearing shales have served as decollement horizons throughout the Valley and Ridge and Allegheny Plateau. The authors believe that processes related to the decollement mechanism can enhance rock porosity and permeability within the decollement zone. Previously it has been concluded that sections of tectonically thickened Devonian strata do exist. Thickening most likely occurs predominantly in shale sections, especially if wedging of competent strata is negligible.

Porosity in tectonically thickened allochthonous shale may be enhanced by the same processes that produced the thickening. Specifically, shale porosity and decollement displacement may be increased by kink band growth in response to an increase in lateral stress. The possibility that locally increased lateral stresses associated with basement faulting may exist has been shown by Advani (1977).[3] That elevated lateral regional stresses existed is demonstrated by Appalachian decollements themselves. It is common knowledge that the flexural slip mechanism is predominant in kink band deformation. Generally the layers remain nearly constant in thickness within, and adjacent to, the kink band while the layers are flexed at the hinge. Kink band growth can increase porosity by three distinct mechanisms.

Figure 7a, (after Ramsay, 1967) illustrates one method of kink band formation. Here, a maximum tectonic compressive stress acts laterally and parallel to layering. The sequence of strata is locally thickened by development of a kink band. During kink band formation a change in area occurs within the sector affected by stresses al and σ_3. This dilation shown by the stippled pattern in the hinge area, produces a tectonically induced porosity (Pi). The length (1) of any layer measured from A to F in Figure 7a, remains the same. The strain produced by this mechanism can be shown as:

$$\text{strain} = \text{A area/original area}$$

Therefore induced porosity, Pi, a direct result of this strain, is:

$$\text{Pi} = \text{area (DCB+FGH) / area AFEB}$$

or (after Ramsay, 1967):

$$\text{Pi} = t/1\ (2 \tan \alpha/2-\alpha)$$

Therefore, Pi is the induced porosity associated with any given kink band, provided the kink band layer parameters do not vary during deformation.

Kink bands can also increase porosity in another manner. Within each kink zone the flexural slip deformation can initiate en echelon tension fractures that cut across kinked strata and are oriented approximately perpendicular to the principal extension direction.

The induced porosity at kink hinges can be graphed for kink bands where $\beta_1=\beta_2$ by plotting Pi with respect to strata inclination'(a) for various t/1 ratios (Figure 7b). A-perusal of the graphed induced porosity values shows that for this type porosity to be significant the t/1 ratio must be greater than approximately 0.5, and should approach 1.0. Admittedly porosity formed by this mechanism may not be large. However, in marginally productive shale wells dependent upon fracture permeability, any induced porosity may be critical. The induced porosity for chevron folds is also plotted at select t/1 ratios for comparison.

Another kink mechanism, that can produce greater induced porosity, occurs where $\beta_1<\beta_2$ (Figure 8a). This mechanism does not strictly entail flexural slip along layers within the kink band. Occurring with flexural slip between bedding outside of the kink band is a component of movement within the kink that is parallel to the principal stress direction. As a result openings form between layer boundaries within the kink band, assuming that layer thicknesses remain unchanged. Here, the induced porosity can be viewed as the strain perpendicular to layering produced by increasing the length of t to t + At. The induced porosity associated with the kink zone, provided At is constant, is given by (after Ramsay, 1967).

$$Pi = (\sin \beta_2/\beta_1) - 1$$

The induced porosity produced by this mechanism is graphed in Figure 8b. Induced porosity is plotted versus $\beta_2-\beta_1$ for various constant values of β_1. Generally $\beta_2-\beta_1$ will not exceed 10^o. Also β_1 should seldom be less than 45^o and should approach 60^o. Within these limits, porosity could be increased by up to 10 percent (0.1 Pi).

Construction of a kink band model for $\beta_2<\beta_1$ would show contraction across layering. This situation is uncommon. However, if bedding contraction does occur porosity can be produced by the formation of en echelon extension fractures or brecciation of the kink band.

The tectonically induced porosities for the kink band models could be lowered by deposition of secondary mineral matter within dilation zones.

The occurrence of kink bands in tectonically thickened decollement shales beneath the plateau would not be surprising. The Burning Springs anticline resembles a box fold (Gwinn, 1964) that possesses all the characteristics of a large conjugate kink (see Faill, 1969). Also, observed outcrops across the flanks of the Warfield anticline show planar limbs and suggest that growth of this structure may be related to a kinking process. Finally, to the east in the folded Plateau, Devonian shales within the Deer Park anticline and other structures are deformed into small upright folds with sharp hinges (Wheeler, personal communication).

Horizontal shortening of at least ten percent has occurred in allocthonous strata across the Burning Springs anticline. It has been demonstrated that a shortening of this magnitude is sufficient for kink band formation. Patterson and Weiss (1966) show the visible inception of kink band formation in deformed laboratory specimens at five percent shortening. They show marked kink band deformation at ten percent shortening that produces β angles approaching 50^o.

Another factor supports the concept of kink development in plateau shales. Ramsay (1967) shows that a given amount of kink band shortening requires less finite shear strain along stratum surfaces than for the same shortening developed with chevron folds. This should also apply to other folds formed associated with tectonic shortening of a stressed column of thin-bedded strata. It follows that the amount of work required to produce kink shortening would also be less.

Pore pressure inherent in decollement horizons can affect kink band development and resulting deformational style. It is known that changes in mechanical properties of rocks undergoing kink deformation affect kink development. Increasing pore pressure should not be uncommon in deforming thin-bedded decollement strata. Evidence that pore pressure did exceed lithostatic pressure within the Devonian shales cored by the Nicholas-Combs well in Perry County, Kentucky, approximately 30 kilometers (19 miles) west of the Pine Mountain overthrust outcrop, is indicated by horizontal layer parallel partings filled with vertical mineral fibers.

Elevated pore pressure would act to decrease ductility and cause rocks to behave in a brittle fashion. If pore pressure increased concomitant with kink development faults may develop along kink zones with resultant brecciation. Faults following kink zones would ideally be inclined approximately 60^o instead of the expected 30^o to the horizontal. High dip of fault planes is suggested by Devonian shale cores from the Nicholas Combs well (Kulander, Dean, Barton, 1977). Here six northwest dipping fault planes occur within 6.7 meters (22 feet) of cored section. Three faults dipped between 45^o and 65^o. Five of the six faults dipped at angles higher than 30^o.

CONCLUSIONS

1. Regional fracture patterns can be separated into domains. Three coal fracture domains have been mapped in southwestern West Virginia. Regional systematic and nonsystematic coal fractures predate Alleghanian folding. These fractures are the result of brittle failure and propagated perpendicular to the greatest tension. Therefore coal fracture domains delineate paleostress domains.

2. Fracture porosity may be increased in fracture domain boundaries or within select fracture domains.

3. At least four different regional principal stress events have affected Paleozoic rocks in southwestern West Virginia from pre-Alleghenian Pennsylvanian time to the present.

4. Regional lateral stresses produced by any fold generating decollement mechanism can be altered over a basin boundary modification such as the Rome trough. The fact that a pre-Alleghenian fracture domain boundary coincides exactly with the underlying Rome trough basement fault zone indicates that regional stresses not related to folding may also be modified by basement structure.

5. Fold geometry on either side of the eastern flank of the Rome trough and Silurian Williamsport folding in the Warfield structure, suggest that slight west to east decollement movement may have occurred into the Rome trough. Eastward movement would be facilitated by an increased gravitational potential produced by uplift of the Cincinnati arch during Carboniferous time. Also the basement gradient steepens along the west flank of the trough.

6. It follows from conclusion five that the limit of western decollement movement in southwestern West Virginia may lie somewhere between the Mann Mountain and Warfield anticlines at one or all decollement levels. Relative Mann Mountain-Warfield fold chronology is important in locating any such western decollement limit.

7. Regional decollement movement and local decollement thickening may be facilitated by formation of kink bands in thin bedded shale units. Kink band development would act to increase rock porosity.

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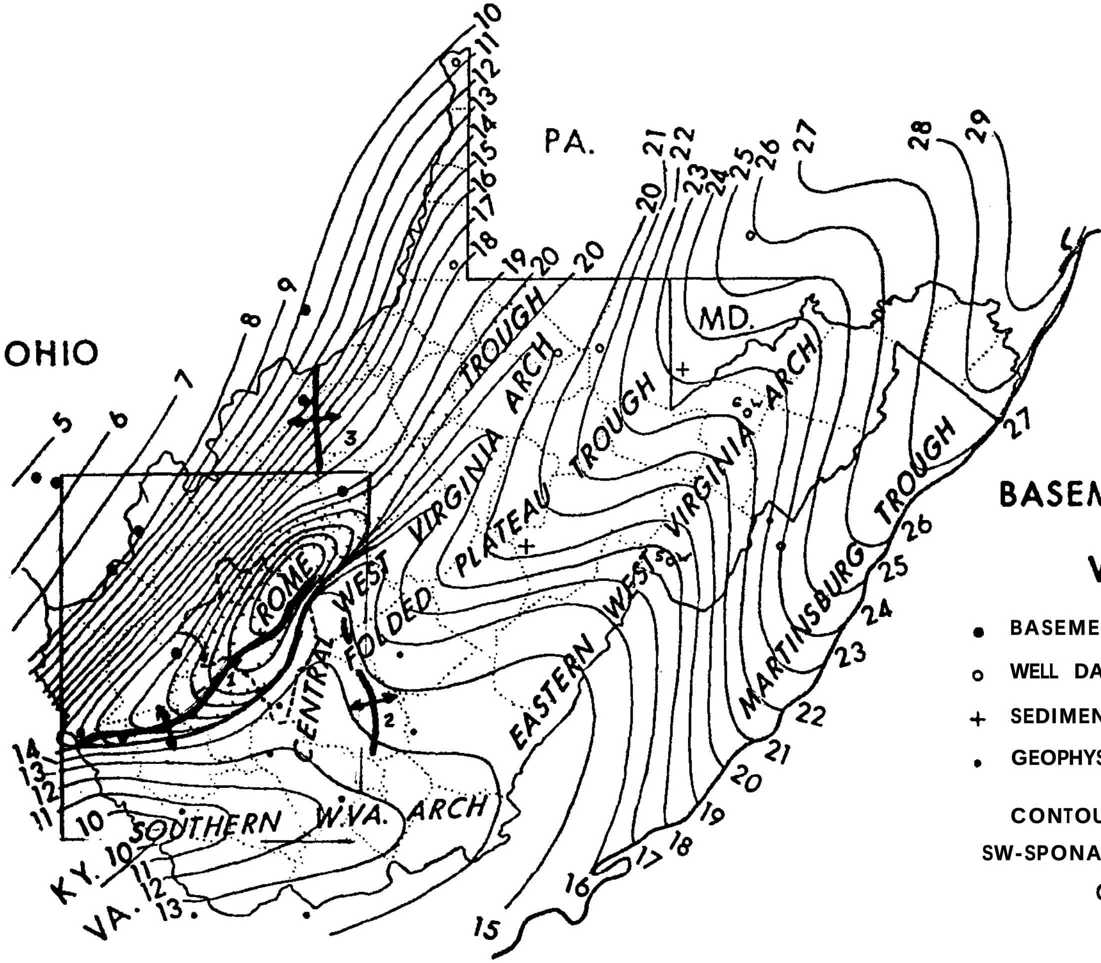

Figure 1

BASEMENT DEPTH MAP OF WEST VIRGINIA

- ● BASEMENT TEST
- o WELL DATA AND SEDIMENTARY PROJECTION
- \+ SEDIMENTARY THICKNESS PROJECTION
- • GEOPHYSICAL CALCULATIONS - MAGNETIC

CONTOUR INTERVAL = 1000 FEET, SUBSEA

SW-SPONAUGLE 'WELL GW-GREENLAND GAP

0 30 60
MILES

Rectangle encloses area under investigation. Anticlinal axes in heavy black;
1) Warfield anticline, 2) Mann Mountain anticline, 3) Burning Springs anticline.
From Kulander and Dean, 1978

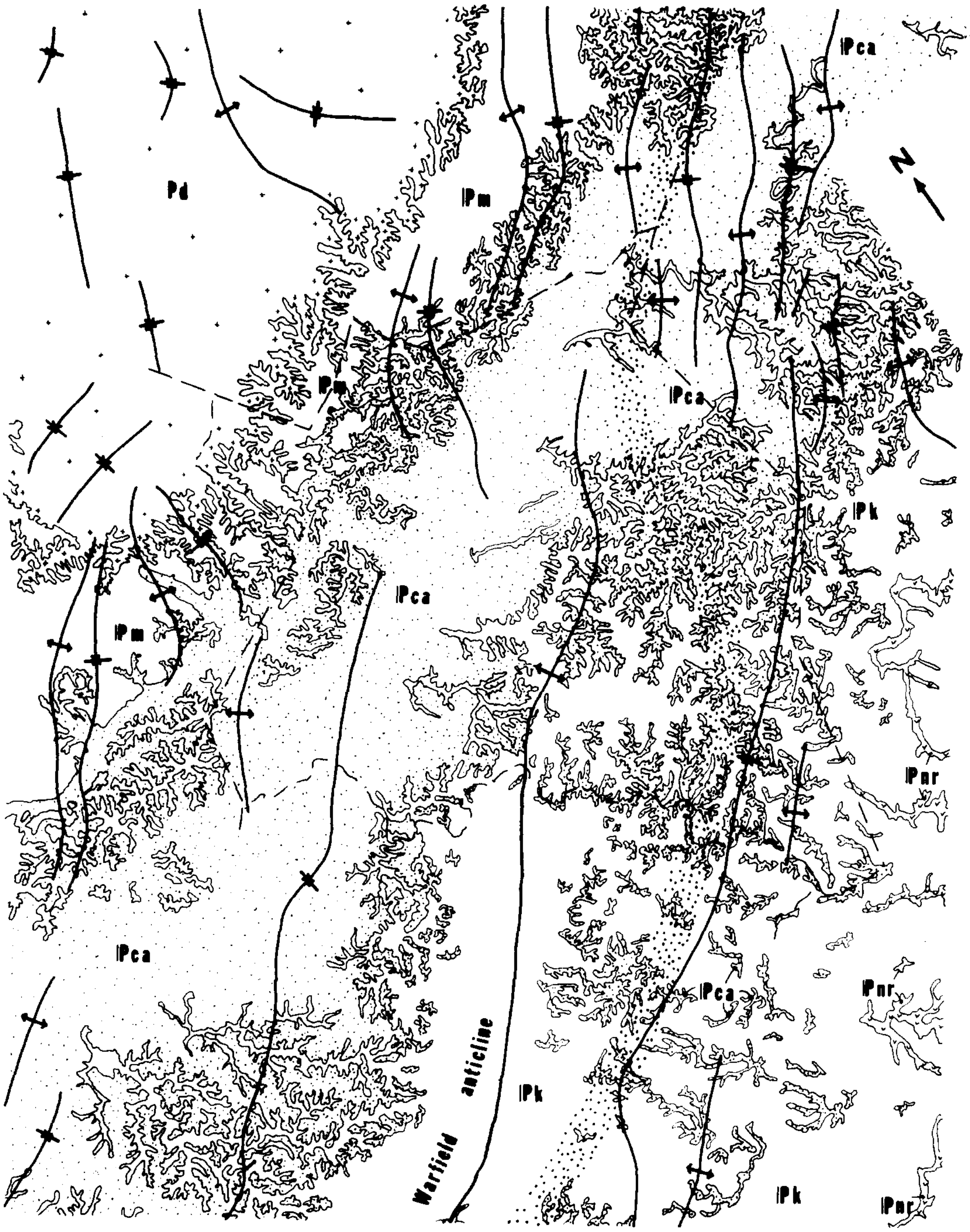

Figure 2, geologic map of Kanawha County (dashed) and surrounding regions. Pd=Dunkard Group (crossed), 1Pm=Monongahela Group (blank), 1Pca=Conemaugh Group and Allegheny Formation (stippled), 1Pk=Kanawha Group (blank), 1Pnr= New River Group (checked). Stippled band indicates zone of subsurface basement faulting. Geology from Geological map of West Virginia, Cardwell, et al., 1968. Scale = 1:560,000.

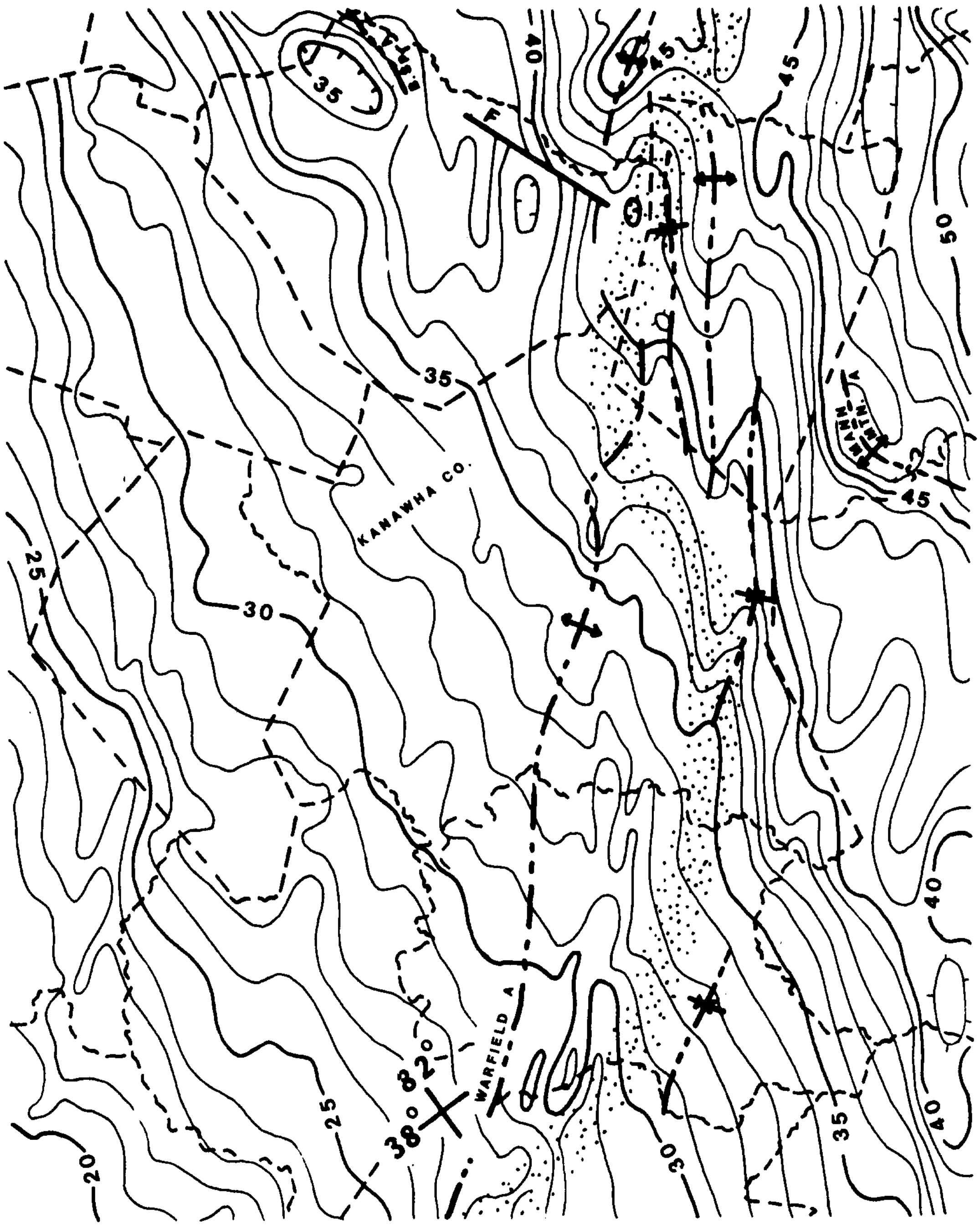

Figure 3, isopach map of middle Devonian through middle Mississippian strata constructed from structure contour maps of the Onondaga (Cardwell, 1973) and Greenbrier (Haught, 1968) limestones. Contour interval in 100s of feet. Scale, 1:600,000

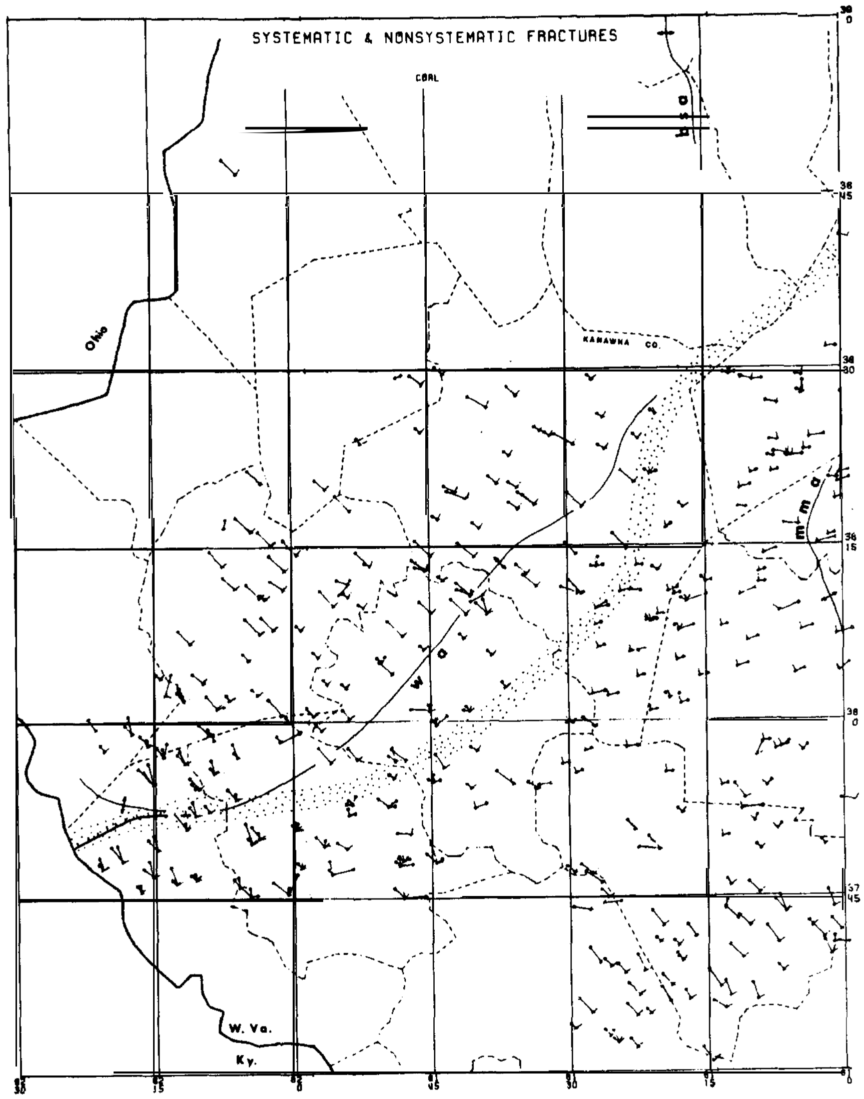

Figure 4, systematic-nonsystematic fracture trends in coal. Barb with ball indicates mean systematic fracture trend. Naked barb indicates mean nonsystematic trend. systematic barb length is inversely proportional to the standard deviation of ten fracture set azimuth bearings taken over each outcrop extent. Systematic-nonsystematic barbs originate at station locations. Stippled band indicates zone of subsurface basement faulting. Scale, 1:1,000,000

Figure 5, tendential view of regional and local systematic-nonsystematic fracture sets. The regional systematic fracture set parallel to arrow and the abutting orthogonal non-systematic fracture set predate folding. A second local systematic fracture set, parallel to bedding slickensides and pen, and related orthogonal non-systematic fracture set, formed after the regional fractures. Local systematic fractures may be a reflection of fold related strain. The abutting relationship can be used to determine fracture chronology. Local fold axes are perpendicular to the pen and local dips are less than 10 degrees.

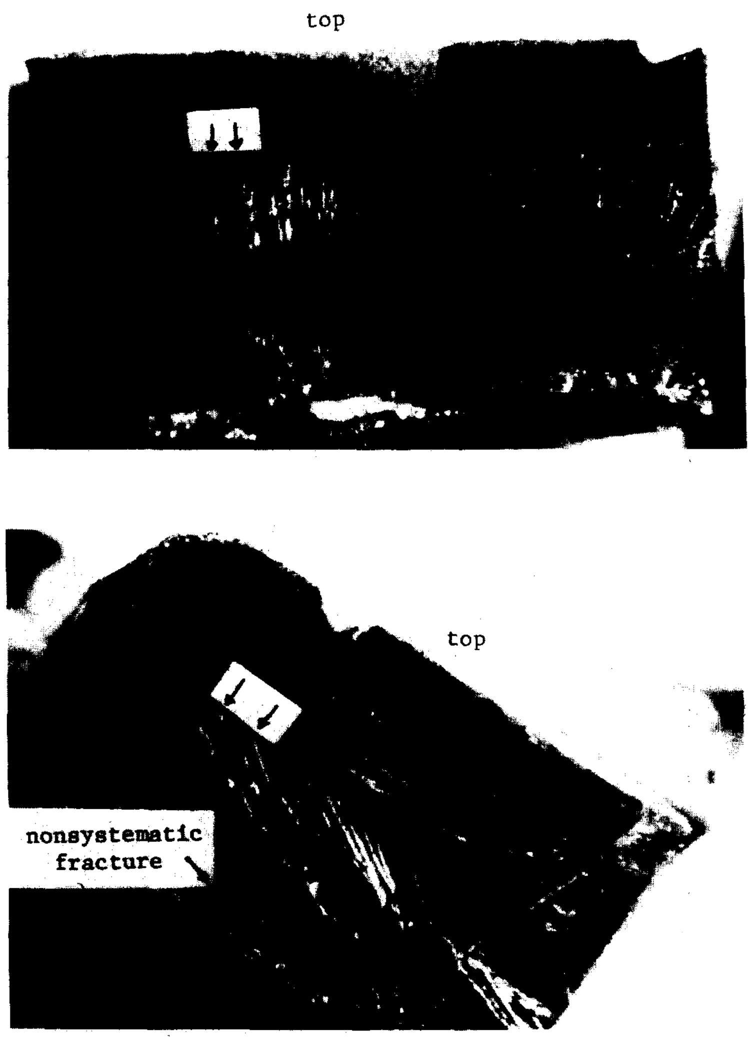

Figure 6, closely spaced twist and inclusion hackle steps (arrow) on coal systematic fracture surfaces indicate coal fractured as a brittle solid. Fractures initiated at'the top of the coal seam and propagated towards its base in response to a maximum stress along the coal's upper surface. Photomicrographs are of fractures in same coal seam nine millimeters thick.

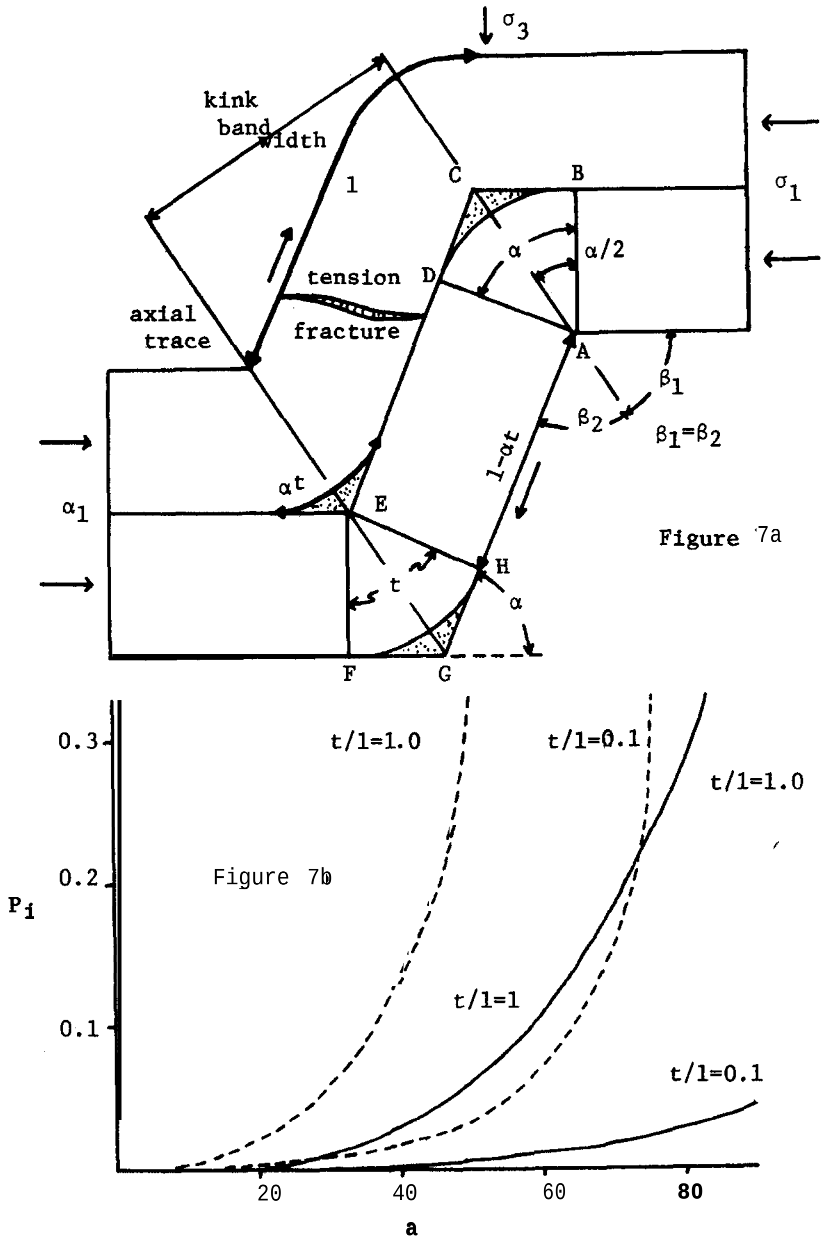

Figure **7a,** idealized kink band morphology where $\beta_1=\beta_2$, and induced porosity is formed at layer boundaries about the axial trace (after **Ramsay,** 1967). Figure 7b shows graphed values of tectonic porosity with variation in t/l ratio and limb inclination a. Pi values (dashed) are graphed for chevron folds for comparison.

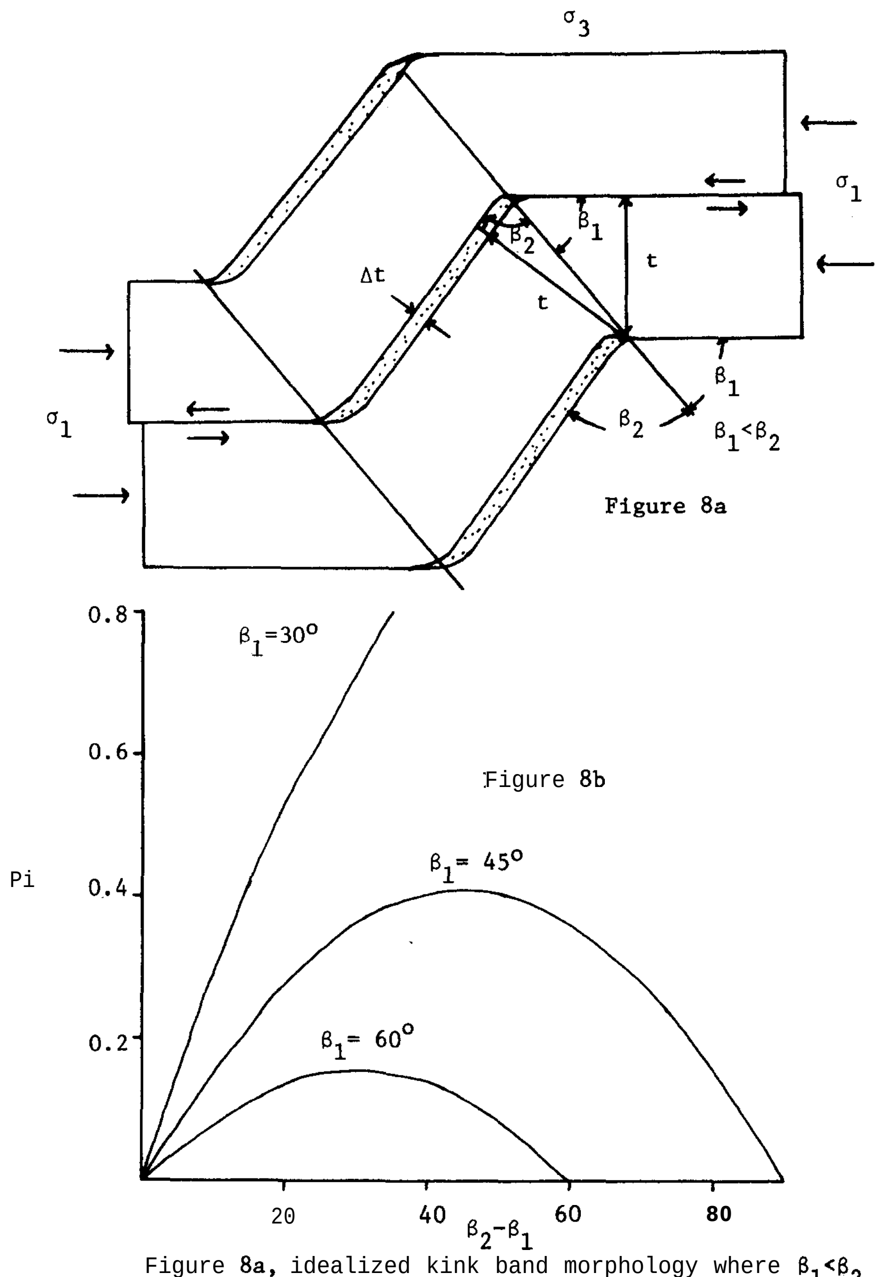

Figure 8a, idealized kink band morphology where $\beta_1 < \beta_2$ and tectonically induced porosity is formed all along layer boundaries between axial traces. Figure 8b, shows graphed values of induced porosity with variation in β_1 and $\beta_2 - \beta_1$.

MANN MOUNTAIN ANTICLINE: WESTERN LIMIT OF DETACHMENT IN SOUTH-CENTRAL WEST VIRGINIA

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ABSTRACT

The Mann Mountain anticline in the Allegheny Plateau of south-central West Virginia is a low, west-facing, thin-skinned fold of aberrant trend with respect to nearly all other Appalachian folds. It and the closely related Ravenseye anticline are rooted in the basal part of the Devonian shale sequence below the Middlesex Shale Member of the Sonyea Formation. The geometry of these folds is directly related to tectonic thickening of the basal Devonian shale units by faulting and possibly by flowage above a decollement at or just above the base of the Marcellus Shale. The Mann Mountain anticline marks the western limit of thin-skinned deformation in this part of West Virginia; its position appears to be controlled by rather abrupt westward thinning of the Marcellus and immediately overlying weak shale beds against a regional unconformity.

Shows of natural gas have been found in the basal Devonian shale interval on the crest of the anticline. Greater fracture porosity and possibly significant amounts of natural gas may be present high on the northwest limb of the anticline at localities where structural closure is greatest or where fracture zones cross the limb.

INTRODUCTION

The Mann Mountain anticline is a low west-facing fold in the Allegheny Plateau province of south-central West Virginia (fig. 1). This fold trends roughly north in southern and central Fayette County to N.26°W. in northern Fayette and southern Nicholas Counties (fig. 2) in contrast to the northeast structural trend of most Appalachian folds. Structural contours on Pennsylvanian coal beds at and near the land surface, suggest that the Mann Mountain anticline plunges north to northwest throughout much of its length. At most, 200 ft (64 m) of closure exists at the surface, restricted to a small area of southeastern Fayette County (fig. 2).

Several deep wells have been drilled along the axis and flanks of the Mann Mountain anticline (fig. 2). From these and other deep wells, Cardwell (1973, 1974) and Perry and Wilson (1977) compiled structure-contoured maps of, respectively, the top of the Onondaga Limestone and equivalent beds (Middle Devonian) and the top of the Oriskany Sandstone (Lower Devonian). Both maps show that the Mann Mountain anticline and the Ravenseye anticline to the east are absent at these depths. Therefore, these anticlines are shallower rooted folds; a decollement exists above the level of the Lower and Middle Devonian carbonate and clastic rocks (fig. 3), which are not involved in the deformation of the shallower cover. These deeper rocks dip gently to the east, southeast, and northeast from central Fayette County (fig. 3). I see no basis for deep-seated faults or folds between existing wells near the Mann Mountain anticline.

Previous speculations concerning the significance of the Mann Mountain anticline include those of Rodgers (1963, p. 1532), who included this fold in a N.20°W.- trending zone of structural disturbance across West Virginia, restricted to the shallow cover above the Salina salt (Upper Silurian). Salina salt is absent in the vicinity of the Mann Mountain fold (Smosna and others, 1977) and, therefore, cannot be a factor here.

Gwinn (1964, p. 890) was the first to suggest that such low anticlines may be rooted in the Devonian shale units. Milici (1980) has provided a regional framework of thin-skinned deformation with respect to oil and gas potential and has delimited the extent of decollement within the Devonian shale units in the Appalachian basin (including the study area).

Structural contours on the Greenbrier Limestone (Upper Mississippian) show that the Mann Mountain anticline involves Mississippian rocks (Perry and Wilson, 1977; fig. 4, present report). Structural closure on the Greenbrier Limestone appears to be roughly the same as that on the Pennsylvanian coal-bearing beds (fig. 2) in spite of the southward and southeastward thickening of the intervening rocks. A simplified "thin-skinned" model of the Mann Mountain fold (fig. 5) prepared for the First Eastern Gas Shales Symposium (Perry and others, 1977), was based on the above data as well as on preliminary correlations within the Devonian shale sequence.

The purpose of the present study is to show, by careful analysis of Upper Devonian and Lower Mississippian rocks, that the Mann Mountain anticline is localized by depositional relationships within the lower part of the Devonian black shale units, specifically, by westward progradation over a regional Middle Devonian unconformity. I will briefly discuss the structural relationships of the Mann Mountain and Ravenseye anticlines with respect to finding natural gas in the Devonian shale.

DISCUSSION

The Greenbrier Limestone (Upper Mississippian) thickens to the southeast and south across south-central West Virginia (Flowers, 1956; Kline, 1976; Perry and Wilson, 1977). The base of the Greenbrier appears to rest unconformably on red beds, sandstone, and siltstone of the Lower Mississippian clastic sequence (fig. 6). Two silty sandstone units, the drillers' "Weir" and "Second Weir," can be correlated from Boone County into southeastern West Virginia beneath the Greenbrier (fig. 6). These sandstone units may represent transgressive marginal marine deposits in a sequence of gray to dark-gray lignitic shale and lenticular sandstone. Correlations based on gamma-ray, density, and other well logs show that a thin, low-density radioactive black shale extends across most of the area about 250-300 feet (75-90 m) below the "Second Weir." This unit is the Sunbury Shale (Lower Mississippian) of the western flank of the Appalachian basin, the "Coffee Shale" of West Virginia drillers. Little if any Berea Sandstone (basal Mississippian) underlies the Sunbury in the area (fig. 6).

In order to prepare an isopach map of the Lower Mississippian clastic sequence, I used the base of the Sunbury to approximate the Mississippian-Devonian contact, realizing that the precise contact may be slightly lower stratigraphically. This clastic sequence thickens rather uniformly southward (fig. 7 and table 1). The purpose of this exercise was to be able to isolate the gross Lower Mississippian sedimentation pattern from the markedly different configuration of the Devonian shale (fig. 8).

The Devonian shale units have a gross depositional trend at nearly 90° to that of Lower Mississippian rocks, particularly in the central part of the area (fig. 8). Westward from the vicinity of the Mann Mountain and Bavenseye folds, the Sunbury interval can be readily determined from gamma-ray logs. The Devonian shale units are anomalously thick under these folds, in direct proportion to the position of the well on the anticline; the structurally higher the well, the thicker the Devonian shale. A comparison of structure at the Oriskany level (fig. 3) below the Devonian shale beds, the structure at the Greenbrier level (fig. 4), and the isopach map of the Lower Mississippian clastic rocks (fig. 7) also leads to the conclusion that the Devonian shale beds must be overthickened under these folds. An additional conclusion, forced by comparison of these maps with the Devonian shale isopach map (fig. 8), is clear: the anomalous thickening of the Devonian shale units under these folds is responsible for the geometry and position of the folds.

The map of Devonian shale thicknesses (fig. 8) suggests that within the Devonian shale units the Mann Mountain fold persists at least 7 miles (11 km) farther south than has been mapped at the surface (fig. 2 and Henry and others, 1977). Alternatively, a structural saddle may exist between the south end of the Mann Mountain anticline and deep well no. 48 in southeastern Raleigh County, as shown on the Greenbrier structure map (fig. 4). Well no. 48 (fig. 8) penetrated more than 4200 feet (1280 m) of Devonian shale, whereas wells to the east and west penetrated less than 4000 ft (1220 m) of shale. If well no. 48 did not penetrate a southern subsurface extension of the Mann Mountain fold, why the anomalous thickness of Devonian shale? Anomalous Devonian shale thicknesses (fig. 8) also extend northeastward from the highest part of the Mann Mountain fold along the axial zone of the Ravenseye anticline and beyond the end of the Bavenseye fold, as mapped on Pennsylvanian coal beds (fig. 2), along the northern extension of the Ravenseye fold suggested by the Greenbrier structure map (fig. 4). Pennsylvanian coal beds (fig. 2) in this area were mapped more than 40 years ago, and the maps may be inaccurate because of 1, topographic inaccuracies in the older base maps; 2, possible miscorrelation of coal beds; and 3, inaccuracies in leveling. In spite of these uncertainties, the previous conclusion holds: the anomalous thickening of the Devonian shale units under the Mann Mountain and Ravenseye anticlines is responsible for the geometry and position of these folds. I conclude that this thickening represents faulting or flowage in the Devonian shale units that can only be attributed to thin-skinned tectonics.

Deep wells in eastern Ranawha, Raleigh, and western to central Fayette Counties show that the Devonian shale units thicken stratigraphically about 30 ft/mi (5.7 m/km) eastward to the western flank of the Mann Mountain anticline (fig. 8), where this thickening rate increases drastically. I have attributed this drastic increase to faulting or flowage. East of the fold's crest, Devonian shale thickness either decreases or remains nearly constant before beginning to increase again in western Greenbrier County. Stratigraphic profile CC' (fig. 9) indicates that much of this eastward thickening is in the shale beds below the Rhinestreet Shale Member of the West Falls Formation of New York (Upper Devonian), in part because of westward progradation (downlap) of Upper Devonian shale units below the Rhinestreet against a major Middle Devonian unconformity documented elsewhere by Wallace and others (1977) and by West (1978).

In addition to this depositional thickening, some tectonic thickening by either faulting or flowage (or both) appears to be present in the pre-Rhinestreet Devonian shale interval in well no. 48 with respect to, wells to the east and west (fig. 8 and 9). Tectonic thickening of the pre-Rhinestreet Devonian shale is even more clearly shown in profile DD' (fig. 10), where reverse faulting of beds below the Middlesex Shale Member of the Sonyea Formation (Upper Devonian) repeats about.100 ft (30 m) of shale in wellno. 20. In the places noted above, we see that the decollement must lie below the Rhinestreet and probably below the Middlesex; from previous work (Cardwell, 1973; Perry and Wilson, 1977), we know that it lies above the Onondaga.

Correlated data (table 2) from geophysical borehole logs in northeastern Fayette County and the adjacent corner of Greenbrier County define more closely the stratigraphic position of this decollement. Correlations within the Devonian shale units are based on West (1978, wells 5 and 6) and standard geophysical well-log correlation techniques. Analysis of these data with respect to wells 13 and 17 (fig. 8), located off-structure east and west of the Mann Mountain fold, indicates that tectonic thickening in intermediate well no. 20 on the crest of the fold is added to an eastward stratigraphic thickening of Devonian shale units of 50 ft/mi (9 m/km) in the vicinity of profile DD', assuming that the off-structure wells contain a normal thickness of Devonian shale. Well no. 28 appears to be on a thickened shale zone under the Ravenseye anticline (figs. 4 and 8). Therefore, structural complications may be present below the Middlesex in this well, at variance with the simple interpretation previously shown (fig. 10). Profile EE' (fig. 11) indicates that this is true. This profile prepared from the data tabulated above (table 2) plus correlations from a gamma ray log of well no. 26, clearly place the decollement in the pre-Middlesex Devonian shale units above the Onondaga Limestone. A close comparison of geophysical borehole logs in this area indicates that the decollement is within the Marcellus-Millboro interval not far above the base of the Devonian shale.

The stratigraphic interval from the top of the Middlesex to the top of the Onondaga decreases steadily westward in off-structure wells (fig. 11), from 387 ft (118 m) in well no. 30, to 241 ft (73 m) in well no. 13, to 149 ft (45 m) in well no. 17 just west of the Mann Mountain fold. The thickness of radioactive black shale units in this interval decreases even more sharply: from 140 ft (43 m) in well 30, westward to 20 ft (6 m) in well 17, a decrease of 86%. Caliper logs from wells on the crest of and east of the Mann Mountain fold show that these shales have sloughed into the borehole, significantly enlarging borehole diameters, whereas caliper logs from wells in the area west of the fold do not show this effect, indicating a westward loss of weak, fractured black shale near the base of the Devonian shale sequence.

I conclude that the Mann Mountain anticline and the western limit of decollement in south-central West Virginia are sited and therefore controlled by the rather abrupt westward thinning of these weak shale units in the basal part of the Devonian shale sequence. As no sufficiently weak shale in the Devonian shale sequence appears to be present west of the Mann Mountain anticline, structural detachment and thin-skinned tectonics did not proceed farther west. The trend of the Mann Mountain anticline and the trend of Devonian shale thickness contours west of the anticline are nearly parallel. I conclude that both the trend and general position of the Mann Mountain anticline are stratigraphically controlled by westward loss of pre-Middlesex structurally weak shales against the regional Middle Devonian unconformity.

SHALLOW GAS POTENTIAL

In western Fayette County and adjacent areas to the north and west, the Greenbrier Limestone (Upper Mississippian) and overlying and underlying sandstones produce natural gas (Kline, 1976; Perry and Wilson, 1977). Paradoxically, farther southeast along the Mann Mountain trend, where the Greenbrier is structurally higher (fig. 4), no commercial gas production from rocks of Mississippian age has been found. Thus, the western flank of the Mann Mountain anticline approximates the eastern limit of Mississippian natural gas in commercial quantities in southern West Virginia. Structurally higher beds of Mississippian age to the southeast are water-filled, and the aquifers are commonly brackish to fresh, indicating recharge or interchange with ground water from the Mississippian outcrop area to the southeast (data in the files of various gas companies.) The presence of water-filled Mississippian rocks on the crest of the Mann Mountain anticline strongly suggests that generation and migration of natural gas into Mississippian reservoir rocks occurred prior to the formation of the Mann Mountain anticline.

DEVONIAN SHALE GAS POTENTIAL

Shows of natural gas in the Devonian shale units have been found on the Mann Mountain anticline. Gas shows in well no. 20 (figs. 10 and 11) were detected from low-amplitude deflections on the temperature log in the vicinity of the fault shown in this well. Several slight gas shows were similarly detected in the basal 130 ft (40 m) of Devonian shale in this well. The borehole was extremely rough and enlarged throughout much of this interval, as indicated by the caliper log, suggesting the presence of fractured shale that had sloughed into the borehole. Similar borehole conditions were found in the Devonian shale interval beneath the Rhinestreet in well no. 48 (fig. 8).

More intense fracture porosity may exist on the western flank of the Mann Mountain fold than on its crest for the following reasons: (1) The Mann Mountain anticline is a west-facing fold, which suggests that the direction of stress release during deformation was to the northwest; (2) extension fractures form more abundantly on steeper limbs (Berger, 1978, p. 72-73). Such open fractures lack slickensides but may be cemented with silica or calcite or may be water-filled. Fracture permeability may be inhibited by slickensides along crestal faults in the Devonian shale on the Mann Mountain fold.

Gas-filled fracture porosity may be enhanced where the amplitude of the fold and consequent dip of the steeper northwest limb is greatest and where lineaments (probable fracture zones) shown on aerial photographs or on Landsat images cross this limb (see Wheeler, in press).

ACKNOWLEDGMENTS

I appreciate helpful discussions with Russell Wheeler and Claude Dean during the preparation of this report. Bill Butler, L. D. Harris, Frank Lesure, John Roen, and Wallace de Witt, Jr., read the manuscript and provided useful criticism and comments. I appreciate the extensive drafting support by Lois Tomlinson. Partial support for this work was provided under Interagency Agreement EX-76-C-01-2287 with DOE Morgantown Energy Technology Center.

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1: present address: U. S. Geological Survey, Box 25046, Mail Stop 934, Denver Federal Center, Denver, Colorado 80225.

2: this volume.

County	Well Permit number	Map number	Depth to base of Greenbrier Limestone (feet)	Depth of base "Sunbury" log marker (feet)	Interval (feet)
Boone	Boo-1021	1	2680	3140	460
Fayette	Fay-41	3	1754	2215	461
	Fay-42	4	2190	2697	507
	Fay-54	5	2140	2586	446
	Fay-60	6	2211	2670	459
	Fay-61	7	2179	2648	469
	Fay-67	8	2555	3060	505
	Fay-69	9	2695	3305	610
	Fay-92	10	2232	2724	492
	Fay-106	11	2776	3368	592
	Fay-112	12	1840	2314	474
	Fay-123	13	2295	2788	493
	Fay-146	14	2642	3184	542
	Fay-174	15	2326	2799	473
	Fay-175	16	2942	2391	551
	Fay-176	17	2272	2821	549
	Fay-197	18	2175	2625	450
	Fay 216	19	2240	2680	440
	Fay-241	20	2309	2798	489
	Fay-243	21	2420	2877	457
	Fay-252	22	2818	3390	572
Greenbrier	Gre-12	23	2071	2610?	539?
	Gre-14	24	568	1320	752
	Gre-15	25	1622	2325	703
	Gre-16	26	2575	3068	493
	Gre-17	27	1230	1818	588
	Gre-18	28	2553	3044	491
	Gre-20	29	1080	1760	680
	Gre-22	30	3294	3820	536
Kanawha	Kan-158	31	2380	2838	458
	Kan-1343	32	1890	2374?	484?
	Kan-1657	33	2914	3337	423
	Km-2481	34	2490	2922	432
Nicholas	Nic-52	37	1728	2162	434
Raleigh	Ral-16	39	3450	4138	688
	Ral-17	40	2496	3004	508
	Ral-36	41	1534	2246?	712?
	Ral-51	42	2835	3505	670
	Ral-83	43	2508		6601
	Ral-86	44	3084	3764	680
	Ral-219	45	3139	3755	616
	Ral-265	46	3173	3900	727
	Ral-289	47	2652	3200	548
	Ral-296	48	3238	3936	698
	Ral-297	49	3045	3792	747
	Ral-336	50	3075	3667	592
	Ral-342	51	2727	3261	534
Summers	Sum-1	54	1980	2777	790
	sum-5	55	2254	3005	751

Table 1. Selected Mississippian well data within the Mann Mountain area (fig. 1). Contacts taken from gamma ray-density logs where possible, based on log correlation network.

Permit number	Well number	S feet(meters)	R feet(meters)	M feet(meters)	0 feet(meters)
Fay-176	17	1089(332)	3986(1215)	4546(1386)	4685(1428)
Fay-241	20	477(145)	3505(1068)	4244(1293)	4729(1441)
Fay-123	13	728(222)	3808(1161)	4532(1381)	4773(1455)
Gre- 18	28	369(112)	3537(1078)	4353(1327)	4832(1473)
Gre- 22	30	342(104)	3494(1065)	4471(1363)	4858(1481)

Table 2.--Depths below sea level of the base of the Sunbury Shale (D), top of Rhinestreet (R), top of Middlesex (M), and top of Onondaga (0) in five critical wells in northeastern Fayette and adjacent Greenbrier Counties, West Virginia.

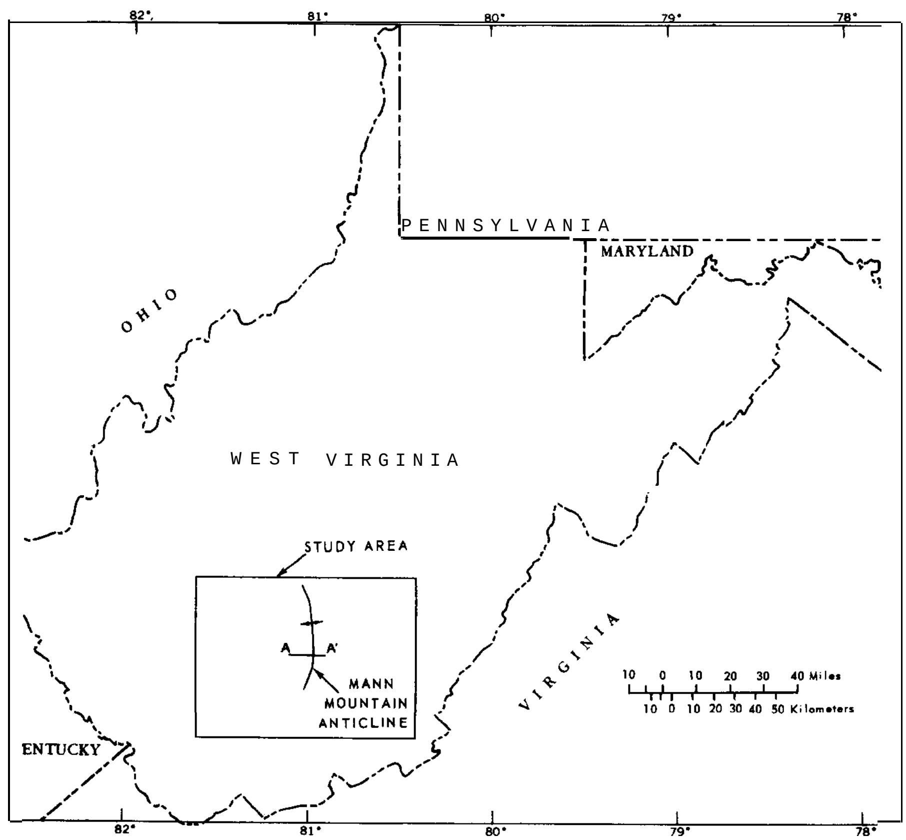

Figure 1- Index map. Box shows area considered in this report. For structural profile AA' across Mann Mountain anticline, see figure 5.

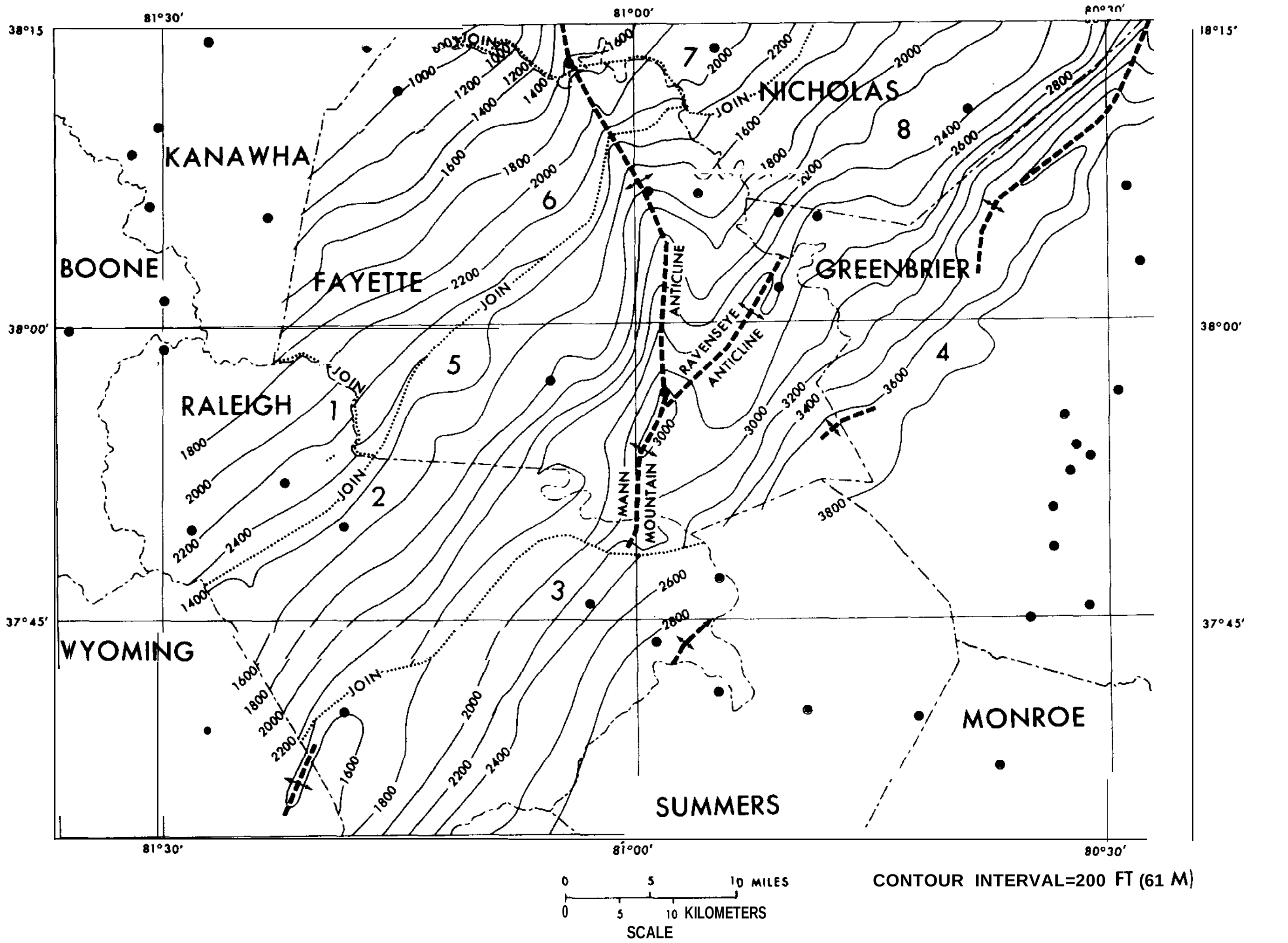

Figure 2– Map showing structural contours on base of several Pennsylvanian coal beds. Compiled from West Virginia county geologic maps. Dotted "Join" lines locate change of datum from one coal bed to the next, necessitated by outcrop and subcrop distribution. Structural contours by the following authors were compiled for numbered subareas: 1, "Eagle" coal bed of Krebs (1916); 2, "Sewell" coal bed of Krebs (1916); 3, "Pocahontas no. 3" coal bed of Krebs (1916); 4, "Sewell" coal bed of Price and Heck (1937); 5, "no. 2 gas" coal bed of Hennen (1919); 6, "Sewell" coal bed of Hennen (1919); 7, "Eagle" coal bed of Reger (1920); 8, "Sewell" coal bed of Reger (1920). Boundaries between subareas 2 and 5, 5 and 4, 5 and 8, and 4 ond 8 ore respectively the Raleigh-Fayette, Fayette-Greenbrier, Fayette-Nicholas, and Greenbrier-Nicholas Counties' mutual boundaries. Solid circles inside open circles represent deep wells penetrating the Oriskany Sandstone (Lower Devonian). Heavy dashed lines represent the axes of anticlines shown on the West Virginia county geologic maps.

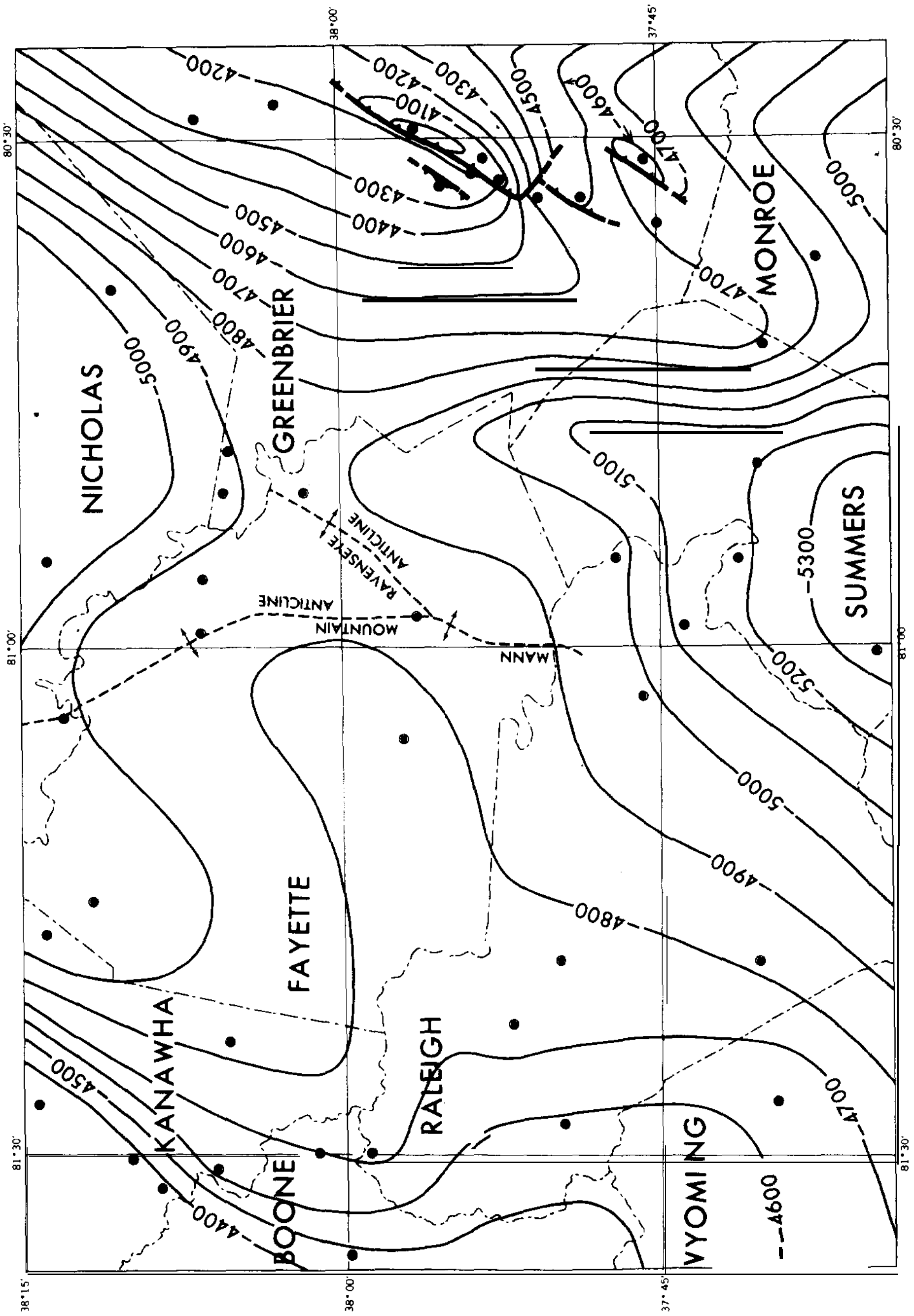

Figure 3– Structural-contour map of top of Oriskany Sandstone (Lower Devonian), south-central West Virginia. Modified from Perry and Wilson (1977). Control wells are shown by deep-well symbol (solid circle inside open circle). Position of axes of Mann Mountain and Ravenseye anticlines from figure 2.

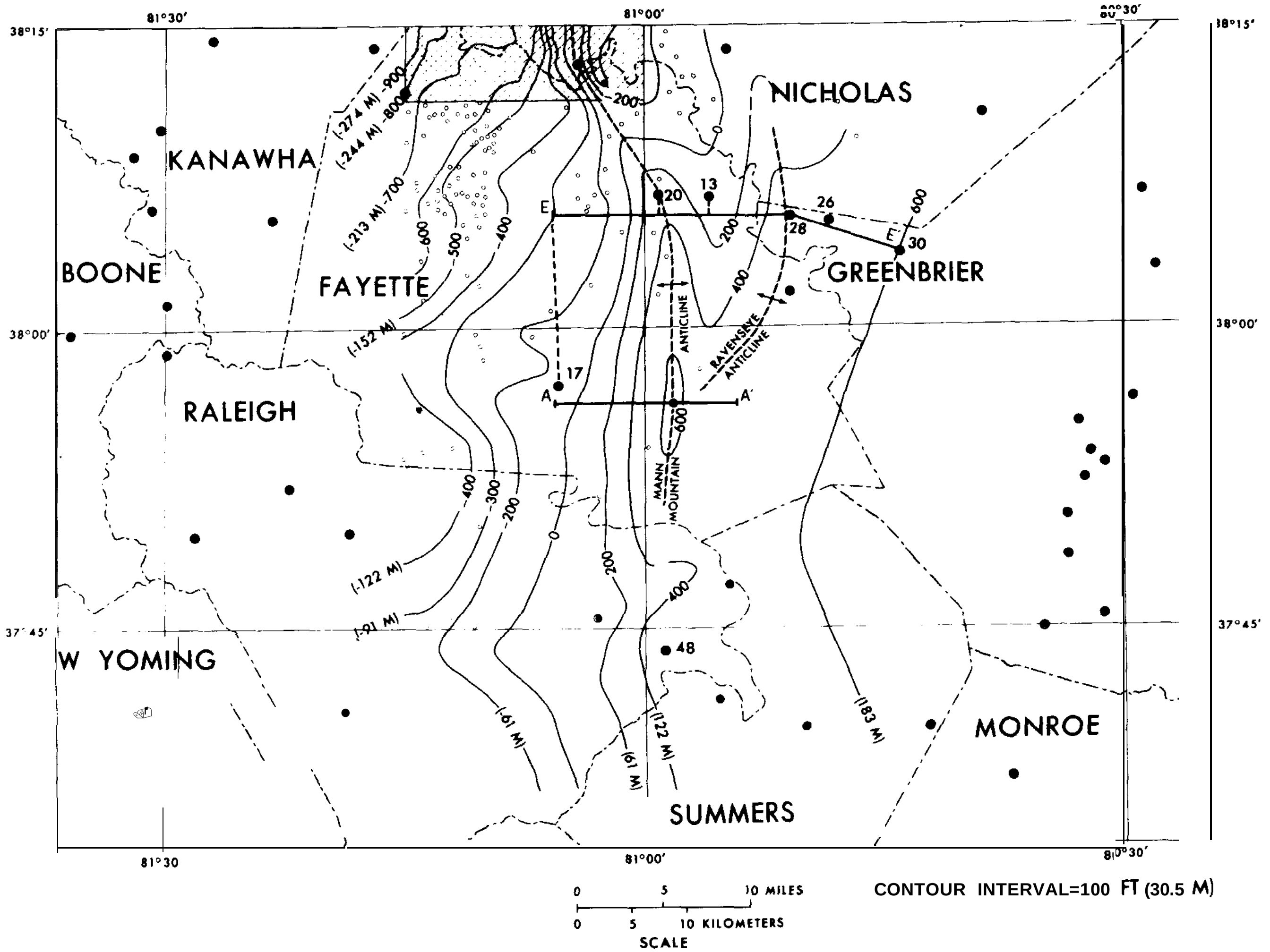

Figure 4– Mop showing structural contours of the top of the Greenbrier Limestone (UPper Mississippian). Modified and expanded from Perry and Wilson (1977); additional data in patterned area (tep center) from Kline (1976, fig. 2). Southern end of Mann Mountain anticline modified from figure 2; based principally on Henry and others (1977) or well or on C. Meissner (unpub. data, 1978), who show that the Mann Mountain anticline does not persist as for southwest os the vicinity of New River gorge. Ravenseye onticline modified from figure 2 on the basis of subsurface data. Small open circles represent selected shallow wells penetrating the Greenbrier. Solid circles inside open circles represent deep wells penetrating Ori skony Sandstone (Lower Devonian). Well numbers ore keyed to table 1. For profiles AA' and EE'; see figures 5 and 11.

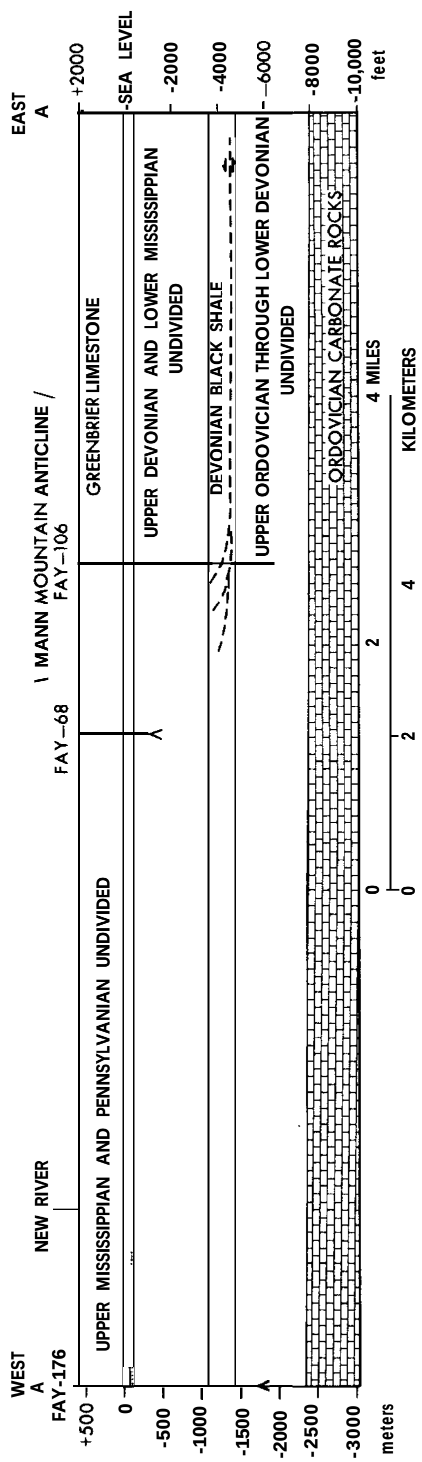

Figure 5— Structural profile AA' across the Mann Mountain anticline in Fayette County, West Virginia. Based on data from Perry and Wilson (1977).

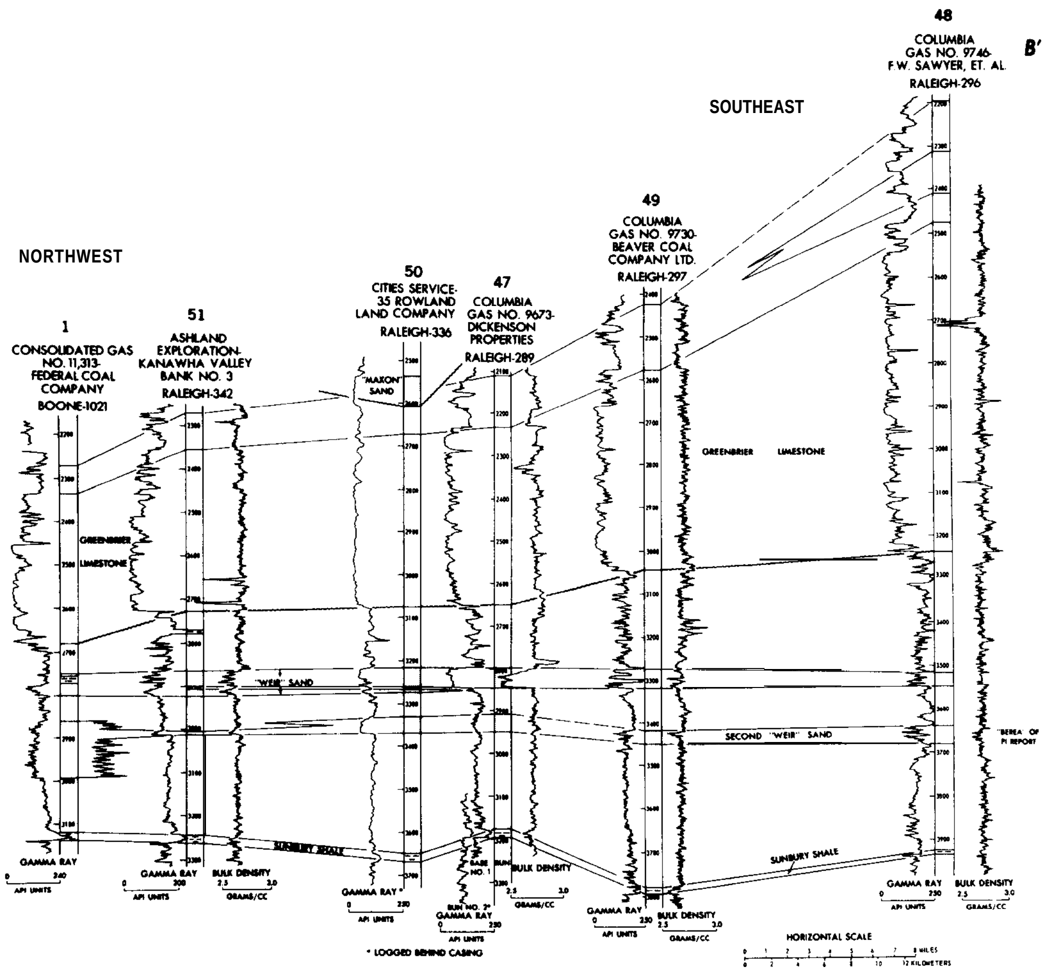

Figure 6— Stratigraphic profile BB' of Greenbrier Limestone (Upper Mississippian) and underlying rocks in south-central West Virginia. Location is shown on figure 7. Well numbers ore keyed to table 1 ond figure 7. Horizontal datum: center of "Weir" sand of drillers. Devonian shale section of well no. 1 correlated by West (1978, well no. 5).

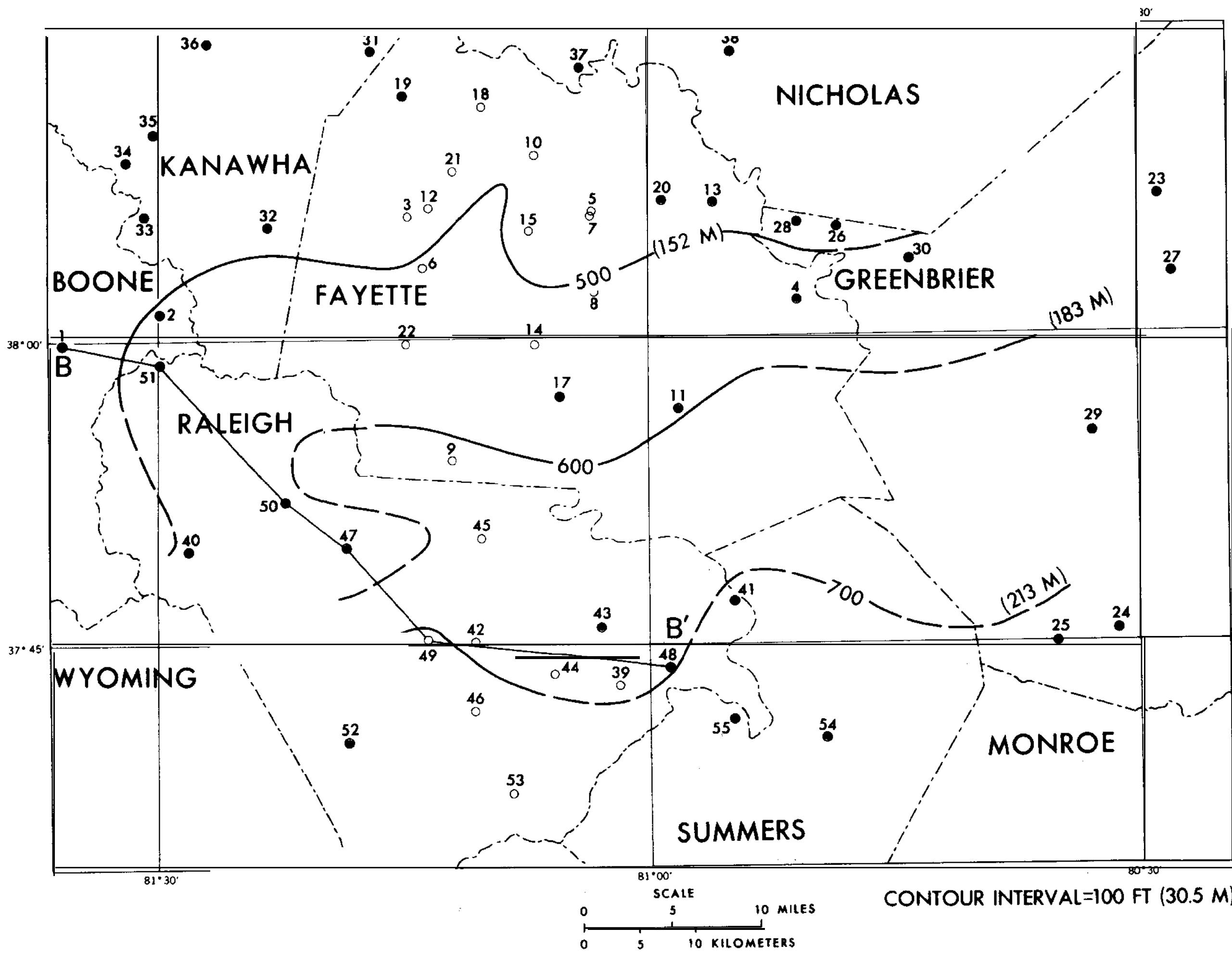

Figure 7– Isopach map of the Lower Mississippian clastic sequence, the interval from base of Greenbrier Limestone to base of Sunbury Shale in south-central West Virginia. Open circles represent shallow wells penetrating the Mississippian-Devonian contact. Solid circles inside open circles represent deep wells penetrating the Oriskany Sandstone (Lower Devonian). Well numbers are keyed to table 1. For section BB', see figure 6.

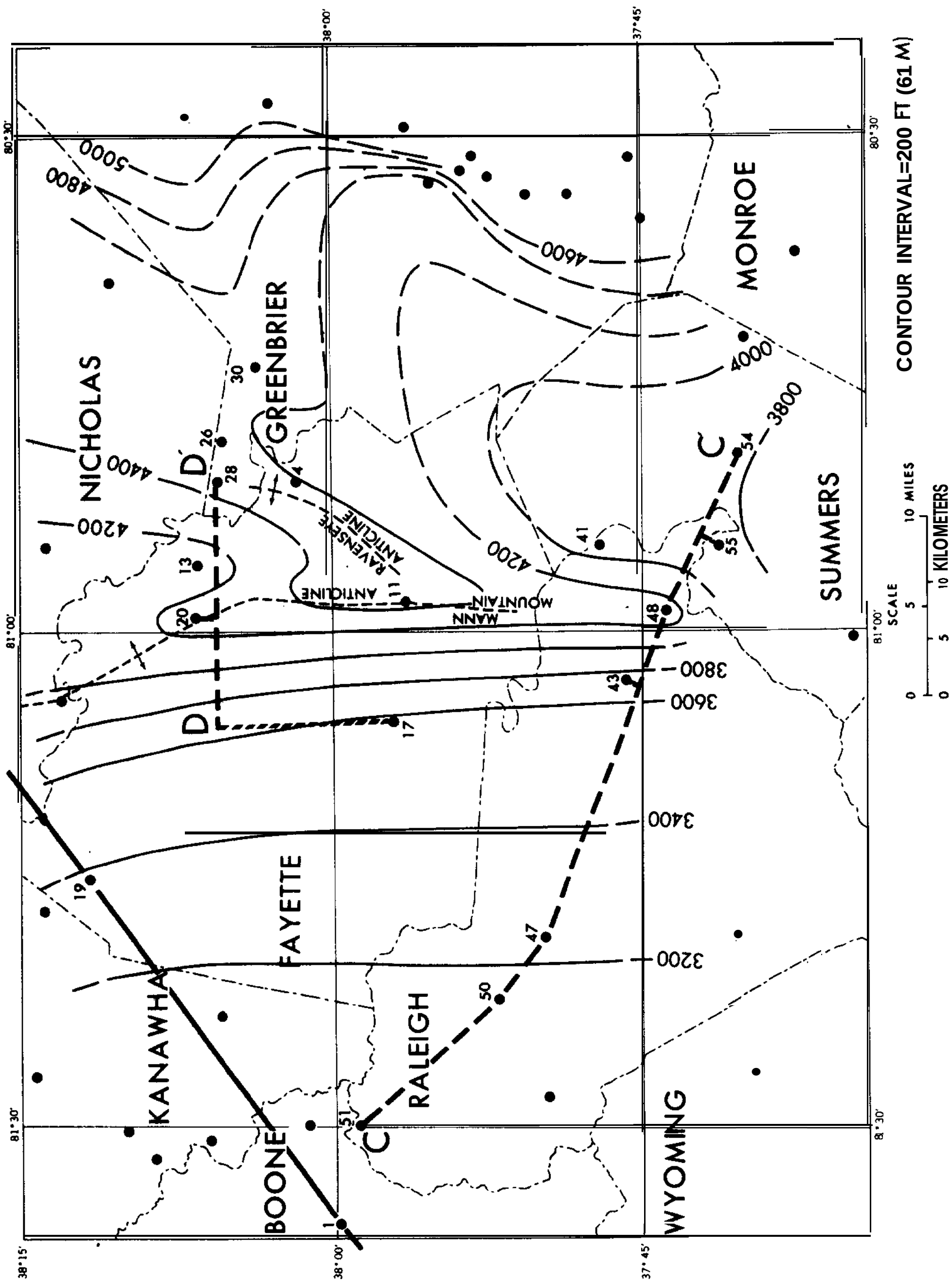

Figure 8—Isopach map of Devonian shale units in south-central West Virginia. Positions of stratigraphic profiles CC' (fig. 9) and DD' (fig. 10) are shown by heavy dashed lines. Heavy solid northeast-trending line ot upper left is the position of the preliminary stratigraphic cross section of Devonian shale units by West (1978) on which Devonian shale correlations in this study are bored. Well symbols ore the same as in figure 7. Well numbers are keyed to table 1.

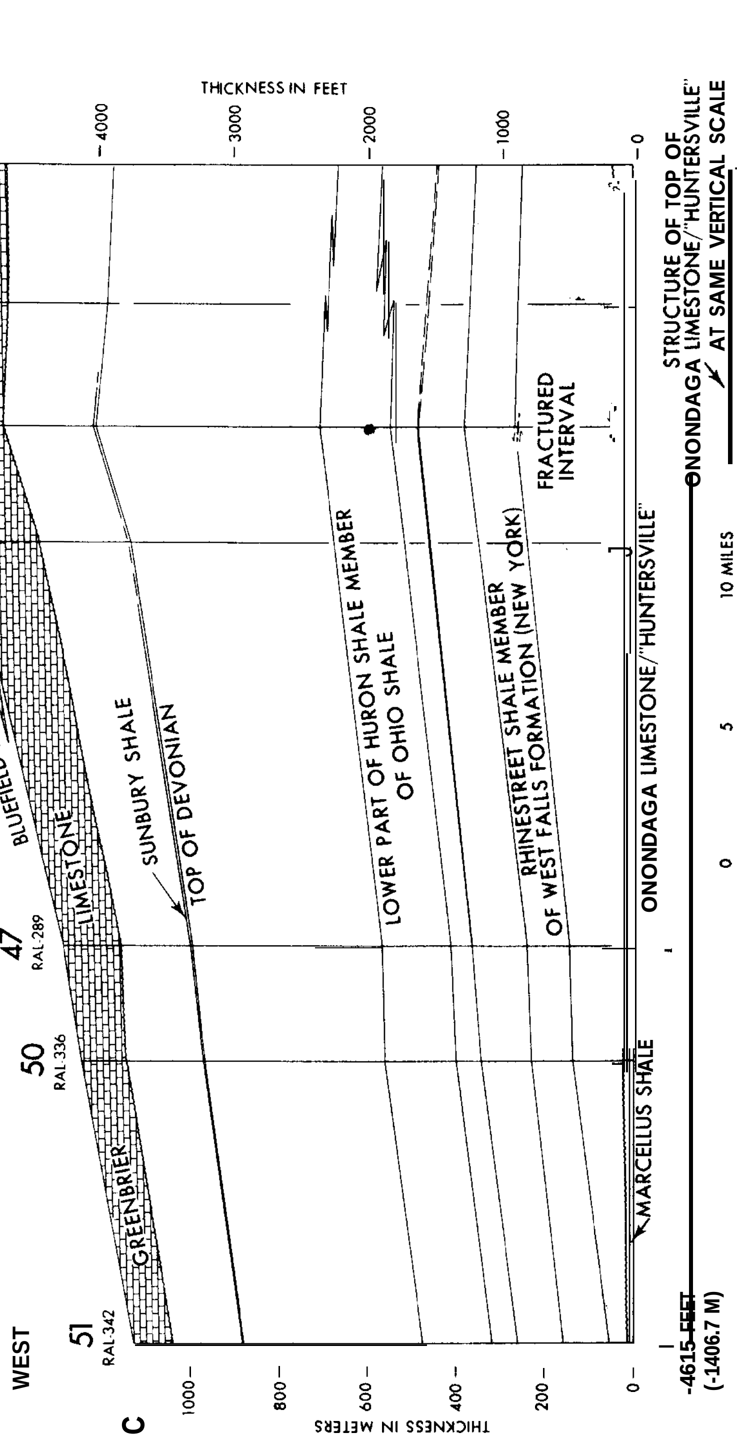

Figure 9.– Stmtigmphic profile of Cc' of Mississippian and Devonian rocks above the Onondaga Limestone (Middle Devonian) or equivalent "Huntersville" Formation and beneath the Bluefield Formation (Upper Mississippian). Devonian shale units correlated from West (1978). Horizontal datum is top of Onondaga Limestone. Location of profile is shown on figure 8. Will numbers are keyed to table 1.

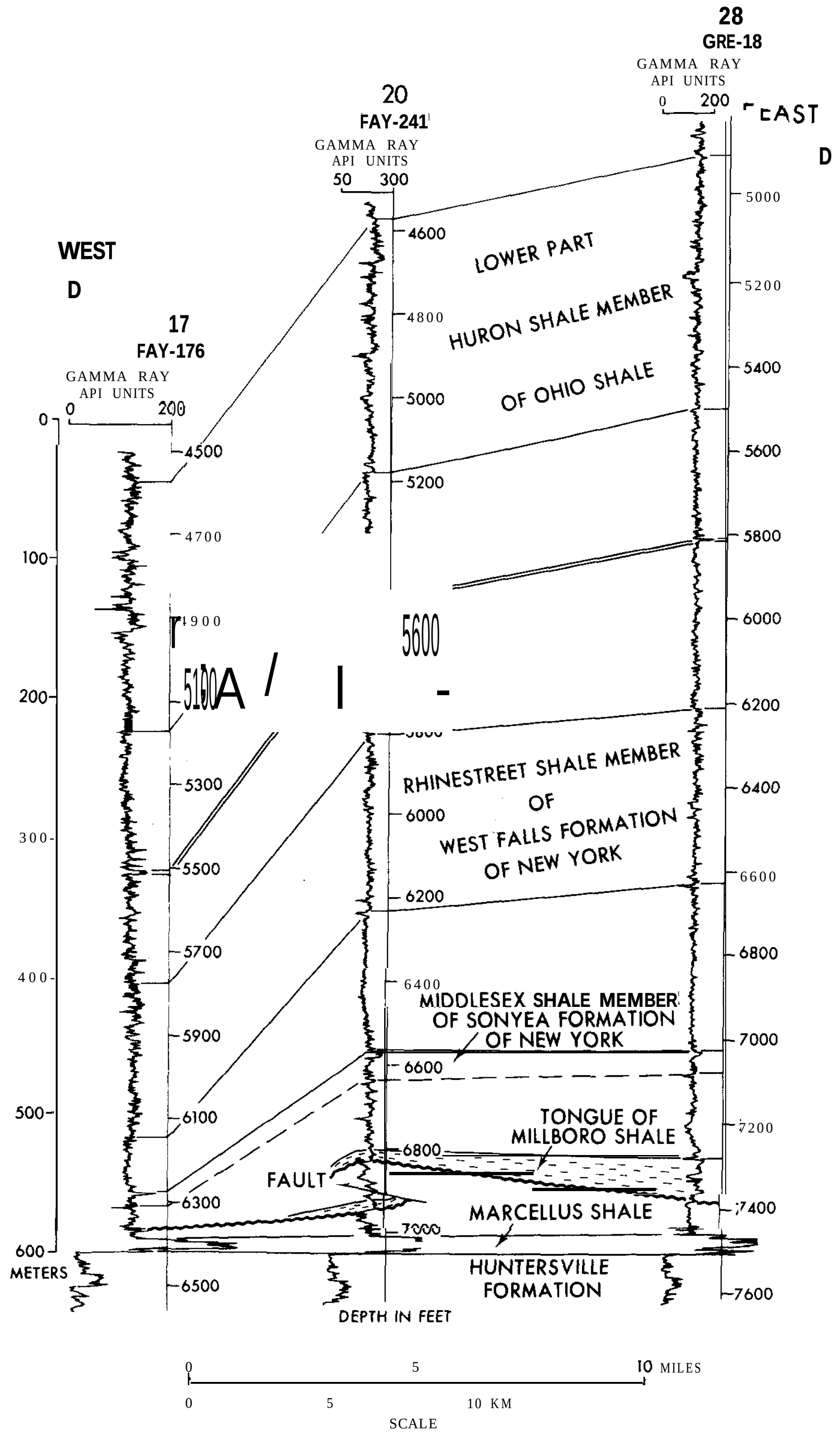

Figure 10–Stratigraphic profile DD' of lower part of Devonian shale units. Devonian shale units correlated from West (1978). Horizontal datum is "Huntersville" Formotion (Middle Devonian) at bore of profile. Location of profile is shown on figure 8. Well numbers are keyed to table 1.

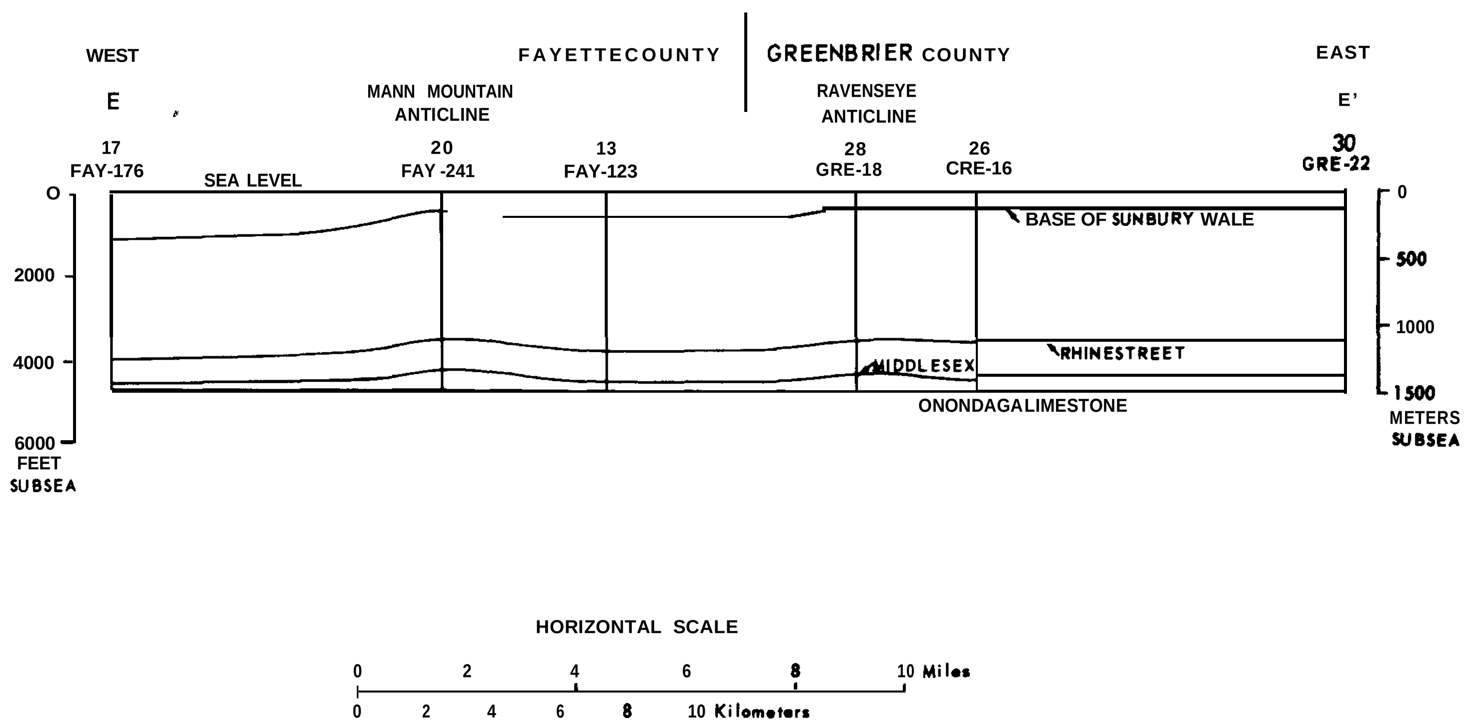

Figure 11 -Structural profile EE' of Mann Mountain and Ravenseye anticlines, Eayette and Greonbrier Counties, West Virginia. Vertical exaggeration: X3.96. Location of profile is shown on figure 4. Well numbers keyed to table 1.

FRACTURE PATTERNS OBSERVED IN CORES FROM THE DEVONIAN SHALE OF THE APPALACHIAN BASIN

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ABSTRACT

This paper examines stratigraphic relationships of core induced and natural fractures and faults in logs and cores from Devonian shales of eastern Kentucky, Ohio, and West Virginia, and suggests models for formation of the observed orientation patterns of slickensides and slickenlines.

Slickensides occur primarily in the more organically rich, less mechanically competent Brown shales, which produce gas in the Appalachian basin. Movement that occurred on these surfaces may have produced differential shortening in the shales (Shumaker, 1978). A stress model (Hafner, 1951) is suggested as the mechanism that produced these slickensides in shales that underwent little or no detachment. The envelope of instability calculated for this model is proposed as the geometrical form of the porous fracture facies.

Slickensides from an interval of the Wise County core #20338 (containing the Pine Mountain thrust) have orientations expected of first and second order faults associated with thrusting (Anderson, 1951). This pattern develops at the top of the Lower Huron shale and thus correlates with the top of the zone of detachment as inferred from contours of blowout zones (Young, 1957). The relationship between core induced and natural fractures, and slickensides, and their restricted stratigraphic occurrence, are also consistent with the concept of the porous fracture facies developed by Shumaker (1978).

INTRODUCTION

This work, motivated by ideas on porous fracture facies (Shumaker, 1978), represents part of research into characteristics of fracturing in the Devonian shale, and their relationship to gas production in these shales. Mechanisms producing these fractures and slickensides are discussed for some of the patterns observed, and their occurrence is predicted to be most intense within the mechanically less competent, organically rich Brown shales. The term "Devonian shale" is used as defined by Patchen (1977) to include all the fine clastic rocks between the top of the lower Middle Devonian Onondaga Limestone and the base of the Lower Mississippian Berea Sandstone. Reference to "Brown sha 1 es" is made for the dark gray to black shale within the Devonian shale, which are usually finer grained, organically rich, and more radioactive than the surrounding shales (Patchen, 1977).

The orientations of core induced fractures and natural fractures including slickensides, and slickenlines observed in cores of the Devonian shale are examined in this report. Unpublished data were collected by the West Virginia Geological and Economic Survey personnel working at the Morgantown Energy Research Center's core lab, and through personal examination of these cores by the authors. Core data investigated in this report were taken from the two Lincoln County (West Virginia) wells #20402 and 20403, the Marietta (Ohio) well #R109, the Martin County (Kentucky) well #20336 and the Wise County (Virginia) well #20338 (see Figure 1). In the Wise County well, the core is taken through the decollement in the Pine Mountain thrust sheet (see cross section Figure 2).

The meanings of the terms core induced fracture, natural fracture, slickenside, and slickenline as used in this paper are briefly discussed here. A core induced fracture contains the fracture origin at the core boundary or on the fracture surface exposed in the core. The fracture surface is marked by divergent hackles that meet the core boundary or pre-existing fracture surface orthogonally (Kulander, Dean, and Barton, 1977). The core induced fractures occur in response to stresses produced by the drilling or core extraction process. Kulander, Dean, and Barton (1977) discuss the characteristics of pre-core and core induced fractures observed in the Nicholas Combs #7239 well, Hazard, Kentucky, and should be referred to for a more complete description. The West Virginia Geological and Economic Survey personnel and the authors of this paper have distinguished between pre-core and core induced fractures using the criteria discussed in Kulander, Dean, and Barton (1977).

A natural (pre-core) fracture is a failure surface formed in response to tensile or shear stresses, and includes unmineralized and mineralized surfaces (some with filled or fibrous void space), and slickenlined surfaces (slickensides). In this study, slickensides are discussed separately, while natural fractures exclusive of slickensides are referred to as "natural fractures." Such natural fractures occur both mineralized and unmineralized, and although not slickenlined, do not necessarily form in response to purely tensile stresses.

RESULTS

Wise County Well #20338

Core induced Fractures:

Initially histograms of strike and dip of the core induced fractures were plotted for each fifty foot interval of the core so that changes in frequency and orientation of induced fractures with depth could be observed. The distributions of strikes (see Figure 3) are approximately unimodal with small variance in all cases except the interval from 5320 feet to 5370 feet where the distribution becomes bimodal with 70" separation between the means of the two preferred orientations. In all eight cases the standard deviation in strike is less than 26°. A visual comparison of these histograms reveals differences between their preferred orientations and suggests the possibility of statistically significant differences between mean strikes,

In general, with circular data of this type, it is not advisable to make linear parametric or nonparametric statistical tests unless the standard deviations of the sample populations are less than 30" (Mardia, 1972). With this criterion satisfied, the Mann-Whitney U test (a nonparametric test of difference between the means of two distributions) was used to determine whether these observed differences are statistically significant (Siegel, 1956). The mean strike of core induced fractures observed in the interval 4870 feet to 4920 feet (Figure 3A) is significantly different at a significance level of $p = .002$ from the mean strike of fractures in the interval 4920 to 4984 feet (Figure 3B), thus representing a significant change in trend of the maximum residual compressive principal stress (sigma 1) of the rock between these two intervals (Overbey, 1969; Komar and others, 1973). The means of histograms B and C are very close so no test was made. There is a significant difference (at $p = .025$) in the mean strikes of histograms C and D. Histogram C is negatively skewed as are A and B, while histogram D is uniformly distributed between N45E and N65E toward the peaked ends of histograms B and C. Histograms D, F, and G have the same mean N56E, while histogram E is clearly bimodal. Since the direction of the maximum compressive residual stress is interpreted as the direction of strike of the core induced fractures (Overbey, 1969; Komar and others, 1973), this bimodality may imply that residual sigma 1 and sigma 2 are approximately equal in magnitude, or that the rock has a fabric producing mechanical anisotropy, or both.

Induced fractures striking NE have bimodally distributed dips with modes at 45NW and 45SE. Similarly, the dips of NW striking induced fractures have modes at- 45NE and 45SW. In intervals where strikes are bimodally distributed, the dips become quadramodal.

The three dimensional orientations of core induced fractures can be summarized using equal area projections of fracture poles. Clustering of poles commonly appears as shown in Figure 4A with the two clusters representing poles to surfaces striking NE and dipping NW and SE. In Figure 4B, there are four clusters representing orientations of poles to surfaces striking NE and NW, and dipping NW and SE, NE and SW, respectively.

Correlation between residual sigma 1 of the rock and the strikes of the core induced fractures permits an examination of variations of residual stress with the stratigraphy of the core. At the top of the core, which is the top of the Cleveland Formation (see Figure 5), residual sigma 1 is oriented N56E (Figure 3A), while below 4920 feet, 30 feet above the top of the Three Lick Bed, sigma 1 swings to N49E (Figure 3B) and continues with this orientation through 4970 feet. The remainder of the Three Lick Bed below 4984 feet, the Upper Huron Shale, and the upper 140 feet of the Middle Huron Shale were not cored, but beginning at 5220 feet in the Middle Huron Shale, the direction of residual sigma 1 is roughly the same as that 236 feet above in the Three Lick Bed (Figures 3B and C). At 5270 feet in the Middle Huron Shale, residual sigma 1 down the remainder of the core does not vary from N56E. Moreover, the dominant preferred orientation of the bimodal distribution is N56E. As mentioned above, the secondary preferred orientation at N14W may correspond to residual sigma 2, fabric, or both. This bimodality becomes most intense in the interval 5312 to 5337 feet in the Middle Huron and corresponds to a relatively more radioactive black shale zone with decreased bulk density.

In addition to changes in the orientation of residual stress in the core, there are also abrupt changes in the density of core induced fractures. The least densely fractured interval occurs from 5370 to 5420 feet (Figure 3F). A more detailed examination of the fracture log reveals that a zone of reduced core induced fracture density occurs between 5356 feet (the top of the Lower Huron Shale) to 5435 feet. The reasons for this decrease of induced fracture density will be discussed below.

Slickensides and Natural Fractures:

In the West Virginia Geological and Economic Survey's fracture log (unpublished), a distinction was made between core induced fractures, slickensides, and natural fractures. Equal area projections of the poles to these surfaces were plotted for each ten feet of the core, and contoured using the Mel 1 is method. A reciprocal relationship was observed between the densities of natural fractures and slickensides, and of core induced fractures, with natural fractures and slickensides increasing in density at the top of the Lower Huron Shale while core induced fracture density decreases. Orientations of natural fractures and slickensides above the Lower Huron Shale (Figure 6) are considerably different from those within it (Figure 7).

In Figure 6A two clusters of poles with average orientations N04E/04NE (a cluster of five) and N81E/15SW (a cluster of nine) correspond to sub-vertical planes striking N86W and N09W respectively. The acute angle of intersection between these two planes is bisected by a line trending N48W corresponding roughly to the direction of movement within the Pine Mountain thrust sheet as inferred from the average trend of slickenlines observed in the thrust (N38W). The significance of this 103" angle between the NE and SW clusters of fracture poles (77° angle between fracture planes) was examined statistically. This has been done by measuring the angles subtended by great circle arcs between the center of the NE set of fracture poles and each of the fracture poles in the SW set. These angles are normally distributed with a mean of 102° and standard deviation of 1.3". The probability that the SW cluster of poles are separated from the center of the NE cluster of poles by 90" or less (determined from the z-value) is less than or equal to 7.02 X 10^{-21}. Similarly, the normally distributed angles between poles in the NE cluster and the center of the SW cluster have a mean of 100° and standard deviation of 2.2", with the probability that the separation is less than 90" being less than or equal to 2.49 X 10^{-5}.

The Coulomb-Mohr failure theory predicts faulting along surfaces forming acute angles with the maximum effective compressive stress (effective sigma 1). If pore pressure is high, tensional effective stresses are possible. If effective sigma 3 is tensional, and effective sigma 1 is less than or equal to the absolute value of effective sigma 3, failure will not only result in shearing between the failure surfaces, but will open them as well, leaving an unslickenlined fracture surface. It is noted that the angle $\theta = 38°$ that these planes make with the inferred effective sigma 1 (Figure 6A) is quite large, with θ around 30" for most rock materials (Hafner, 1951) but as high as 40" for "moderately ductile . . . materials" (Handin, 1966, p. 230).

Another cluster of five poles can be seen at the center of the projection (Figure 6A). These fractures are filled and some contain vertical fibers, and probably represent fractures developed during conditions of high fluid overpressure. Horizontal slickensides are also found In this interval of the core (Figure 6B), and indicate that high fluid overpressure may have been local in nature or that natural fractures and slickenslides developed at different times. The cluster of four subhorizontal natural fracture poles (Figure 6A) trending roughly N20W may have formed under conditions similar to those described in the previous paragraph, but under more brittle conditions (higher pore pressure? shallower depth?) because θ is smaller.

Poles to natural fractures found in the Lower Huron Shale (Figure 7A) form clusters in areas of the projection corresponding to planes striking N60E and N32 W, and exhibit a considerable amount of scatter. The orientations are those expected for longitudinal and transverse joints respectively. The E-W clusters of poles may correspond to a diagonal set of joints.

Poles to slickensides found in the Lower Huron Shale (Figure 7B) form a girdle across the equal area project ion. The poles forming this girdle have an average trend of N38W. From NW to SE across this girdle, clusters plunge 27NW, 55NW, 90°, 77SE, and 62SE. Anderson (1951) suggests that as a major fault develops, stress trajectories adjust to orientations more nearly perpendicular and parallel to the fault plane. This redistributed stress field may in turn produce second order faulting about first order faults (Ramsay, 1967, p. 284). Figure 8 illustrates the relationship between first order faul ts, the re-oriented stress trajectories, and thelr associated second order faults. The horizontal faults are most numerous, and appear in the center of the projection as a cluster of vertical poles. The secondary faults C and C' appear on the projection as poles plunging into the SE octant. The similarity in orientation between the surfaces C and C' may produce some overlap between their plotted poles so that only a single large cluster may be observed. Second order faults A and B will have poles plunging into the NW octant of the projection. The pattern of slickensides shown in Figure 7B indicates that the interval of detachment in the Pine Mountain thrust begins with the top of the Lower Huron Shale and is contained within this shale since no slickensides are observed from 5456 feet to the bottom of the core at 5484 feet. Contours of blowout zones in the area (Young, 1857) put the top of the interval of detachment at about 50 feet above the top of the Lower Huron Shale.

Anderson's (1951) model has been shown by Chinnery (1966) to be in error as a model for strike-slip faulting. Chinnery points out that Anderson's model is limited in that it is based on the assumptions that there is no vertical movement of the ground surface, and that the str ike-sl ip fault extends to an infinite depth. These assumptions allowed Anderson to calculate the stress field (and thus the fault patterns) associated with strike-slip faulting as a two-dimensional problem. The similarity of fault orientations predicted from this model to the pattern of slickensides observed in the interval of detachment in the Pine Mountain thrust indicates that the assumptions made for strike-slip faulting are more accurately made for horizontal thrust faults. Specifically, it can be assumed that there is little or no horizontal movement of material perpendicular to the direction of net displacement of the thrust sheet, and also that the state of stress in any vertical plane through the main body of the thrust fault (Figure 1) parallel to the direction of net displacement is the same for all such cross sections.

The interpretation of the patterns observed on the equal area projections assumes equal weights for surfaces of various inclinations. However, with these observations taken from a near vertical core four inches in diameter, the number of surfaces of a certain orientation intersected by the core will be a function of their density, total surface area, perpendicular spacing and inclination to the core. If all variables except inclination are held constant the probability of intersecting a vertical surface is much smaller than that for intersecting a horizontal surface. For instance, if the perpendicular spacing between all surfaces, regardless of their inclinations, is one foot, with total surface area constant, the number of vertical surfaces intersected by a vertical core four inches in diameter will be one third the number of horizontal surfaces intersected. Taking this into account will have the effect of increasing the number of horizontal poles by some weighting factor, but will not change the locations of clusters of maximum density, leaving an identical pattern of clusters on the projection.

Martin County Well 120336

Core Induced Fractures:

Equal area projections of the poles to induced fracture surfaces were plotted and contoured for each 50 foot interval of the core. The 982 foot core contained 1650 induced fractures that were logged by

West Virginia Geological and Economic Survey personnel (unpublished data). An examination of these projections reveals little variation in the orientations of the core induced fractures throughout the length of the core. In general, the poles to these fractures produce clusters similar to those observed in the Wise County well. The projections are characterized by two clusters which correspond to planes striking generally N50E, and dipping 30NW and 30SE. A notable exception to this pattern occurs in the interval 2500 to 2550 feet in the lower Cleveland Member and the upper Chagrin Shale. This interval is characterized by considerable scatter of poles into the NE and SW quadrants of the net (Figure 9). The gamma ray log indicates that this interval is higher in radioactivity and thus more organically rich than the surrounding shales. No slickensides and only one natural fracture were present in this interval, Three horizontal induced fractures were logged elsewhere in the core and may be natural fractures. Horizontal induced fractures have not been logged for any of the other wells investigated, and, with these three exceptions, none have been recorded to dip less than 20°. In some cases it may be difficult to distinguish between pre-core and core induced fractures. As mentioned earlier, Survey personnel distinguished between pre-core and core induced fractures using criteria established by Kul'ander, Dean, and Barton (1977); however, in many cases, the surface features on the fractures may be indistinct, making identification difficult. The absence of uniformly oriented fractures in this radioactive organic shale may indicate that many of the scattered fractures have been misidentified, and may in fact be natural fractures. Similar scatter found 'In the undifferentiated fractures of the Jackson County (West Virginia) well has previously been interpreted (Shumaker, 1977, oral communication) as due to natural fractures.

Slickensides and Natural Fractures:

Poles to the natural fractures logged from this core are shown plotted and contoured in Figure 10A. Clusters of poles on the edge of the projection correspond to vertically dipping fractures striking about N46W and N36E. These strikes are respectively perpendicular and parallel to the regional trend of the low amplitude folds observed on a coal form line map of the area (Shumaker, 1974). These two fracture sets may represent systematic joint sets associated with these folds. The NE striking fractures are most abundant in the top half of the core, with the NW striking fractures most abundant in the bottom half of the core. If the shales of this core are considered as a relatively homogeneous plate of material, then buckling of this plate by NW-SE oriented compressive stress is likely to produce NE striking longitudinal joints above a neutral surface in the plate, with NW striking cross joints most likely to develop beneath this neutral surface.

Mellis contours of lines representing the trends and plunges of constructed and measured slickenlines are shown for this well in Figure 108. Slickenside strikes and dips were taken from the West Virginia Geological and Economic Survey's fracture log, and measurements of slickenline trend and plunge were made at the University of Kentucky's Core Library in Lexington by the senior author. Personal examination has shown that with few exceptions, slickenlines from all wells trend perpendicular to the strike of the slickensides on which the slickensides occur. Thus the probable orientation that slickenlines would have on a slickenside can be inferred from the strike and dip of the slickenside. This conversion was made on the Survey's slickenside data for the projection shown in Figure 10B. In this figure, the cluster of ten lines in the SE quadrant is roughly centered at N09W/39SE. The horizontal to subhorizontal lines in the NW and NE quadrants are scattered and roughly centered. at N04W/34NW. Hafner (1951) has calculated that for a supplementary horizontal stress field that decreases exponentially with distance from the region of maximum pressure (Figure 11A) potential faults will develop that dip toward and away from the region of maximum pressure (Figure 11B). Slickenlines plunging SE in this model would be more concentrated than those plunging to the NW (Figure 11C) in agreement with slickenline orientations observed in this core (Figure 10B). The variation of stress with depth along the SE side of this block would be due to the Pine Mountain thrust (Figure 11A) as it splays to the surface south of the Martin County well and SE of the Nicholas Combs #7239 well (Figure 1). The zone of instability or zone of potential faulting is shown in Figure 11B. The distance which this zone extends into the Devonian shales beyond the thrust will vary with the composition and thickness of these shales. Layer anisotropy within the shales will also affect the shape and extent of this zone. Although resultant shear surfaces can act to decrease porosity of the shales (Bagnal1 and Ryan, 1976) potential tension fractures associated with shearing might increase porosity and permeability. This zone of instability is suggested as the model for the porous fracture facies of Shumaker (1978). The location of the Martin County well within the facies is indicated by the line AB in Figure 11B. The well's location is inferred from an absence of a continuous girdle of slickenlines through the SE quadrant into the NW quadrant of the projection (Figure 10B) as suggested in Figure 11C for the model. The lack of NW trending slickenlines in the lower part of the Nicholas Combs #7239 well (Kulander, Dean, and Barton, 1977) indicates that that well is located at CD within the fracture facies (Figure 11B). Although the Nicholas

Combs #7239 well is closer than the Martin County well to the Pine Mountain thrust, the Devonian shale sequence is thinner in the Nicholas Combs #7239 well and may restrict the distance which the fracture facies extends beyond the Pine Mountain thrust.

In Figure 12, the trends of slickenlines are plotted with depth in the core. Absolute plunge is shown with the amount or degree of plunge increasing to the right. In the upper part of the core slickenlines trend N40-45W. Beginning with the Lower Huron Shale, slickenlines become more numerous, more variable in trend, and trend more northerly than those above. Examination of the gamma ray log from this core shows that the slickenlines always occur in the Brown Shales and that zones showing the greatest scatter in the trends of slickenlines correspond to relatively more radioactive shales.

Marietta Well #R109

The Marietta (Ohio) well #R109 is located in Washington County (Figure 1). Orientations of fractures were measured by West Virginia Geological and Economic Survey personnel from the 150 feet of core taken from this well (unpublished data), but no distinction was made between core induced and natural induced and natural fractures. Poles to these undifferentiated fractures form clusters in equal area projection similar to those formed by core induced fractures in the Wise County and Martin County wells. The fractures were grouped into 10 foot intervals to examine variations in their orientation with depth. The well was cored from 3500 to 3650 feet in the Huron Shales (which extend from 3164 to 3731 feet) . .

In Figure 13, the contoured equal area projections of the poles to fracture surfaces from each 10 foot interval are outlined. The average strikes of the fracture sets from these intervals ranged from N35E to N78E with standard deviations ranging from 4 to 7 degrees. No statistical comparisons were made between the distributions of strike for each interval, but in several instances, variations in strike appear real, indicating vertical changes in the orientation of residual sigma 1 in the core. Similar variations were discussed earlier for the induced fractures from the Wise County well and also have been observed in the undifferentiated fractures logged by Survey personnel in the Nicholas Combs #7239 well, Perry County, Kentucky (Shumaker, 1977). Scatter observed in the poles to these fractures may represent incipient clusters in the NE and SW quadrants of the equal area net similar to clusters of core induced fractures observed in the Wise County well. The patterns formed by these poles on the equal area projection indicate to us that most of the fractures logged in this core were probably core induced. Although some of the scattered poles may be due to natural fractures, the scatter observed in plots of core induced fractures in the Wise County #20338 well does not permit differentiation between pre-core and core induced fractures on the basis of orientation alone.

Lincoln County Wells #20402 and #20403

The cores from the Lincoln County wells #20402 and #20403 in southwestern West Virginia (Figure 1) were examined for slickensides. In the #20402 well most slickensides occur in the sixty-four foot interval between 3894 feet and 3958 feet. In the #20403 well the slickensides occur mostly in the 267 foot interval between 3766 feet and 4033 feet (Figure 14). In both wells, the upper part of the cored interval is the Angola member of the West Falls Formation and the lower part is the Rhinestreet member of the West Falls Formation and unconformably the underlying Hamilton Group. These slickensides occur in the organic rich, high gamma radioactive Brown shales of the Devonian shale.

In the #20402 well the slickensides are well developed, covering the entire diameter of the surface (Figure 15). They are slightly less well developed in the #20403 well. Slickensides range from smooth and well polished to rough and uneven; none of the surfaces is mineralized.

The trends and plunges of twenty-two slickenlines in well #20402 and twenty in well #20403 were measured from an oriented core by goniometer and Brunton compass. In each case the slickenlines were perpendicular to the strike of the slickenside.

The dominant set of slickenlines in well #20402 trends N47W, and plunges about 25" either to the northwest or to the southeast (Figure 16A). A minor set trends N30°E and plunges 30" to the northeast and southwest. Neither set is restricted to a particular depth interval or a particular lithology within the interval analyzed. However, the slickensides of the minor set tend to be rougher than those of the dominant set. No slickenside in the minor set is smooth or polished as are many of those in the dominant set.

The dominant trend of the slickensides in well #20403 is centered at N20°W (Figure 16B). The average plunge of the slickenlines is about 30" to the northwest or southeast. The slickensides are not as well developed as those in the #20402 well, The preferred orientations of the slickenlines in the two wells suggest a northwest trending movement, probably tectonic in origin, as opposed to more random orientations that might be related to slumping and compaction. The opposing plunges of the similarly oriented surfaces can be explained as differential movement in incompetent beds,

CONCLUSIONS

We have examined the stratigraphic distribution of slickensides and other natural fractures observed in five cores in the Devonian shales, and have found that these failure surfaces are confined almost entirely to the darker, organically rich, highly radioactive Brown Shales within the sequence. These 'observations indicate that adjustments to stresses applied to the Devonian shales occurred primarily within these Brown Shales, even for relatively intense stress concentrations as in the Pine Mountain thrust. These observations support the concept of the porous fracture facies discussed by Shumaker (1978).

The general NW trends of slickenlines observed in the Lincoln and Martin County cores, and the Nicholas Combs #7239 core, suggest that they are related to regional tectonics and may represent differential shortening in the basal shales (Shumaker, 1978). NW trending slickenlines observed in the Brown Shales of the Nicholas Combs #7239 core (Kulander, Dean, and Barton, 1977) and slickenlines observed in the Martin County core are explained by the stress model proposed by Hafner (1951). The zone of instability predicted for this model is suggested as the model for the porous fracture facies discussed by Shumaker (1978). The similarity in orientations of natural fractures in the Wise County core through the Pine Mountain thrust with those from the Martin County core indicates that both groups of natural fractures represent adjustments to similar stress fields.

In the Wise County core, the intensely fractured, organically rich, highly radioactive, Lower Huron Shale corresponds to the interval of detachment in the Pine Mountain thrust. The model of first and second order faul t i ng (Anderson, 1951) accounts for the orientations of slickensides observed in this interval.

ACKNOWLEDGEMENTS

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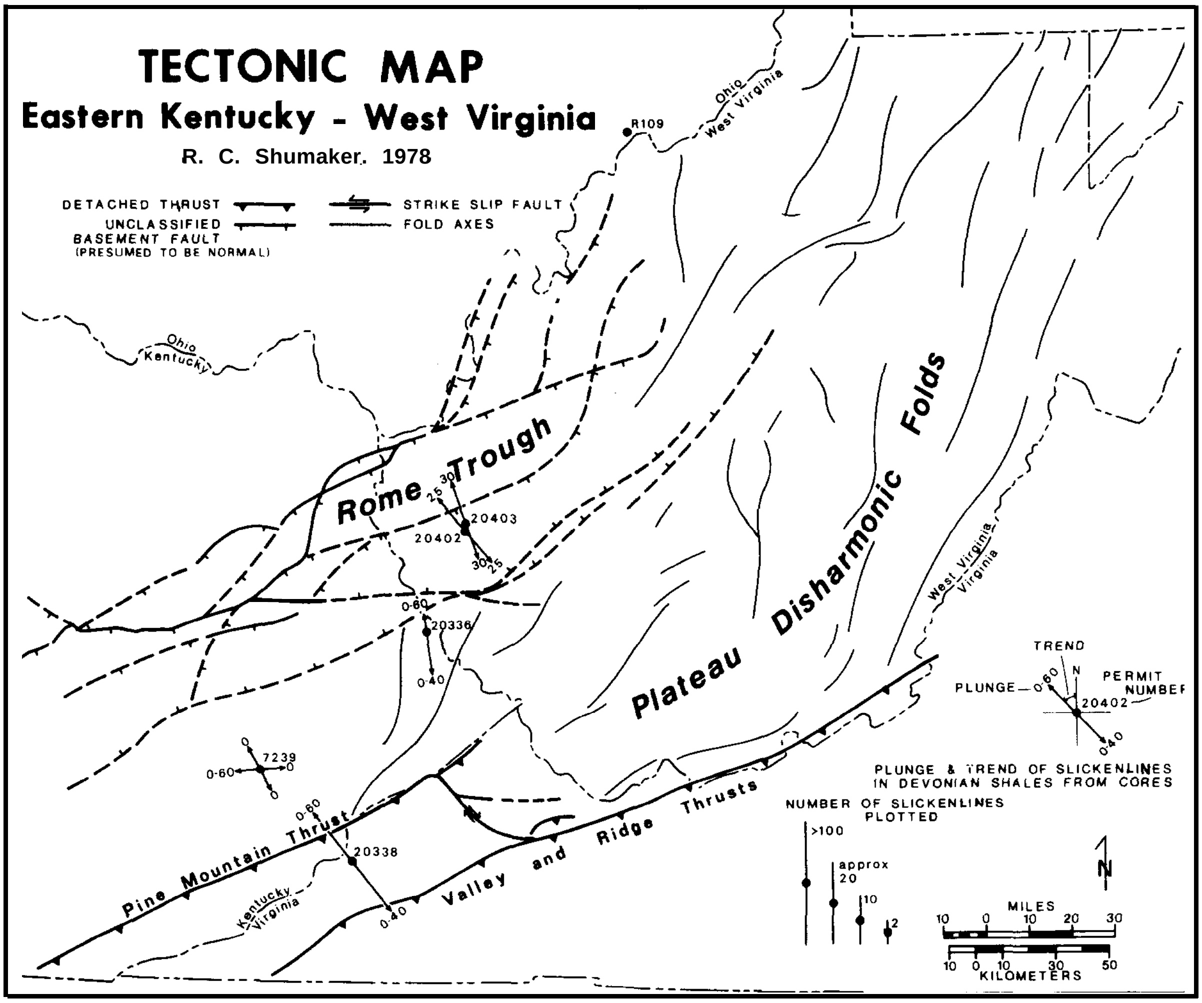


Figure 1 The cored wells examined in this paper are designated by their permit number. Dominant trends and plunges of slickenlines are indicated for these wells.

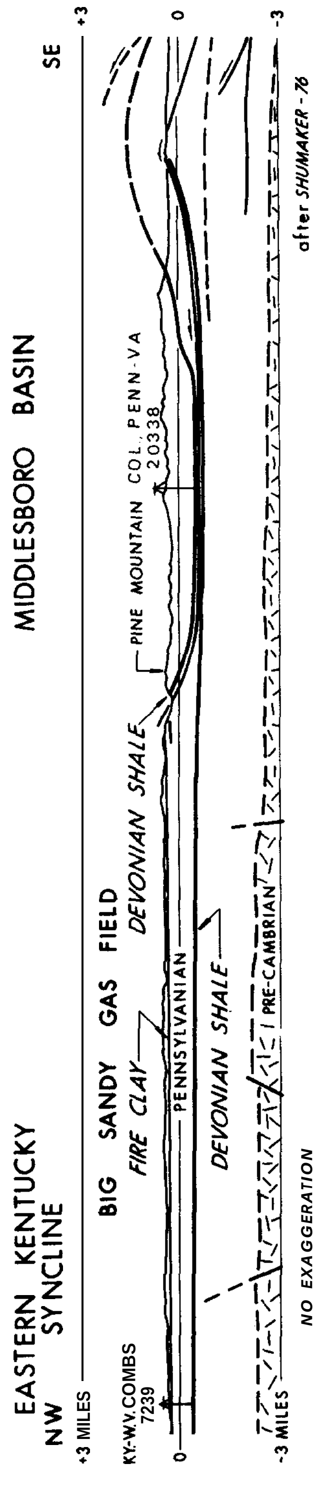
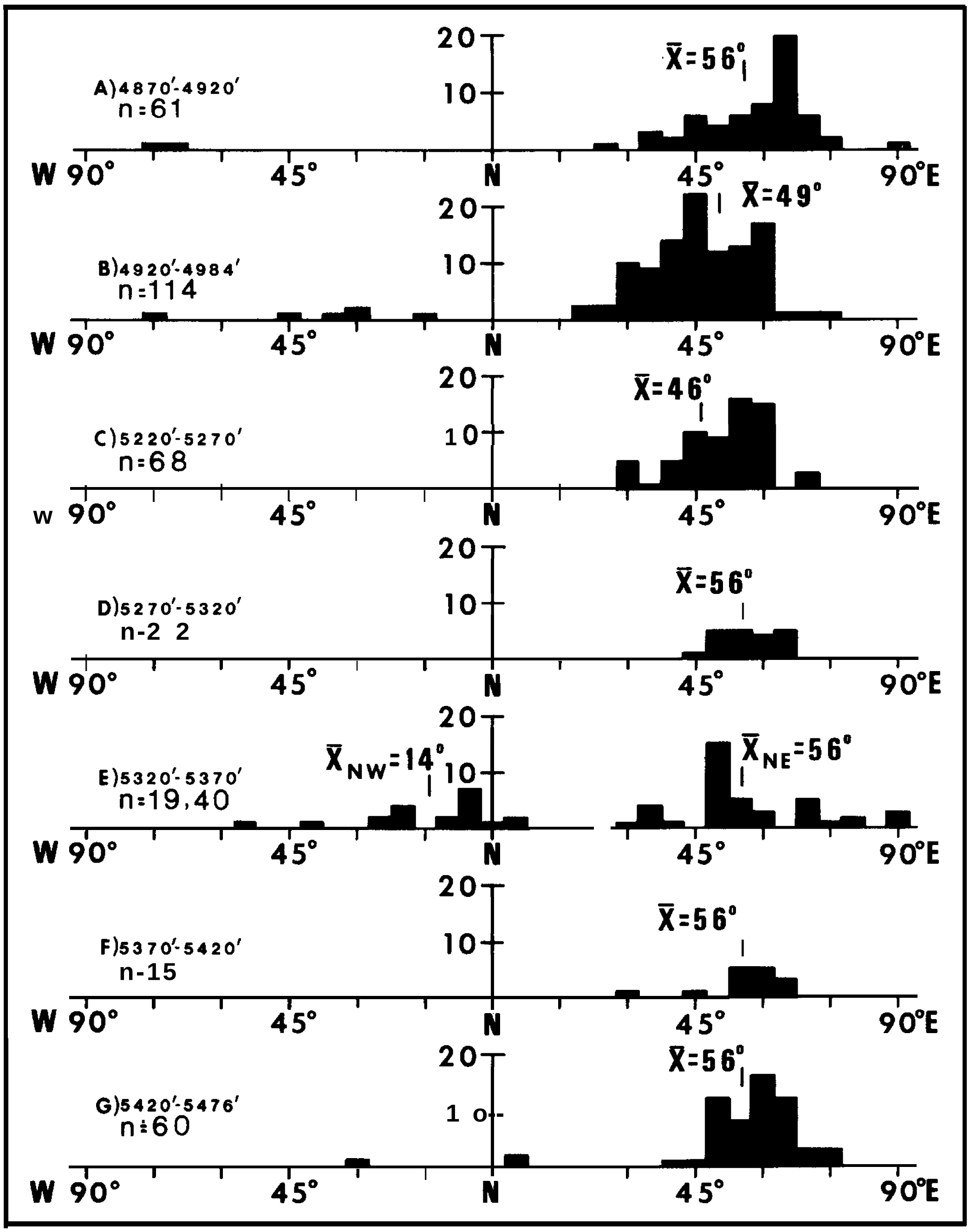


Figure 2 This cross section through the Pine Mountain thrust shows the locations of the Nicholas Combs #7239 (Perry Co., Ky.) and the Wise County #20338 (Virginia) wells.



FREQUENCY HISTOGRAMS: STRIKES OF INDUCED FRACTURES PLOTTED IN 5° - 50 FOOT INTERVALS.

Figure 3

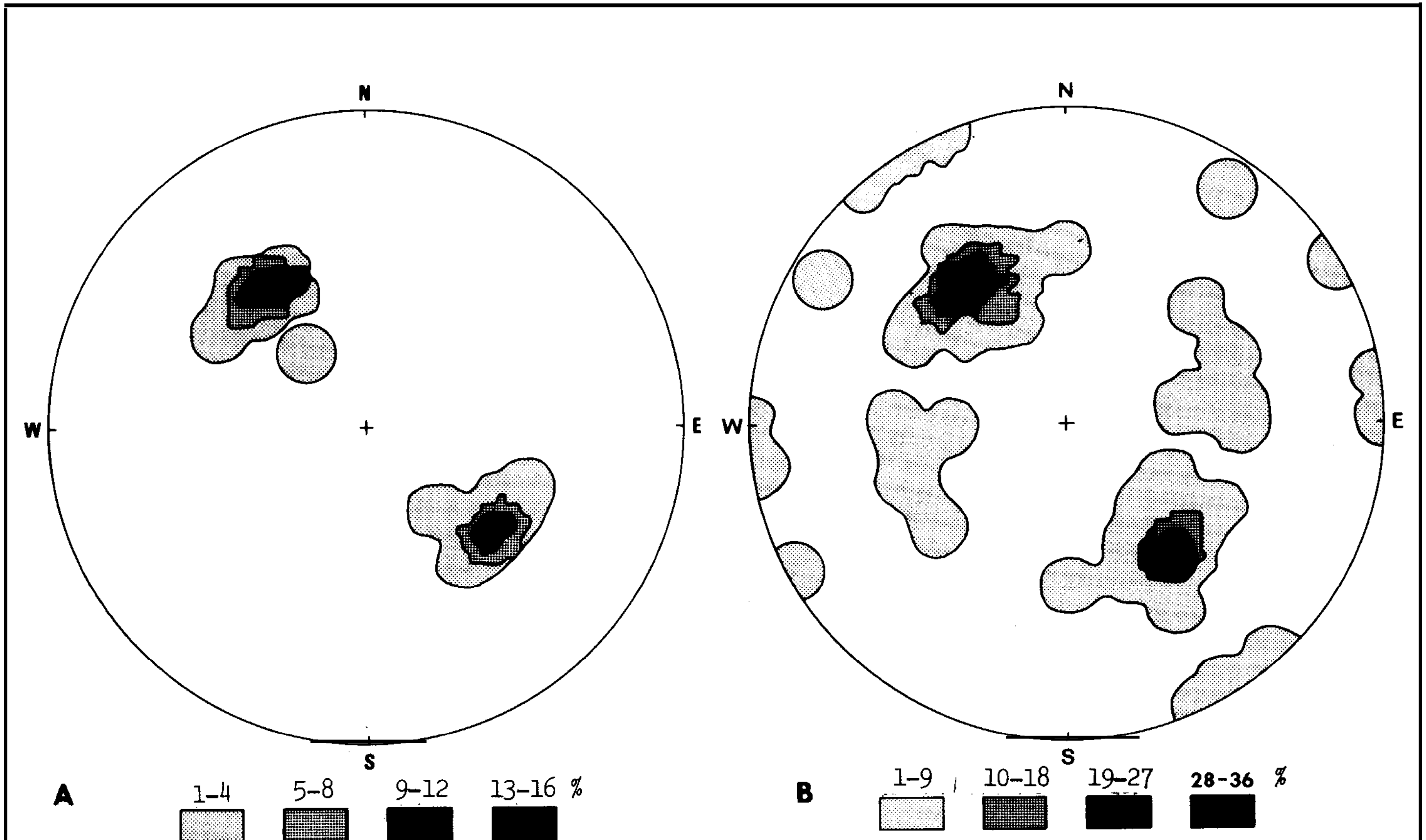
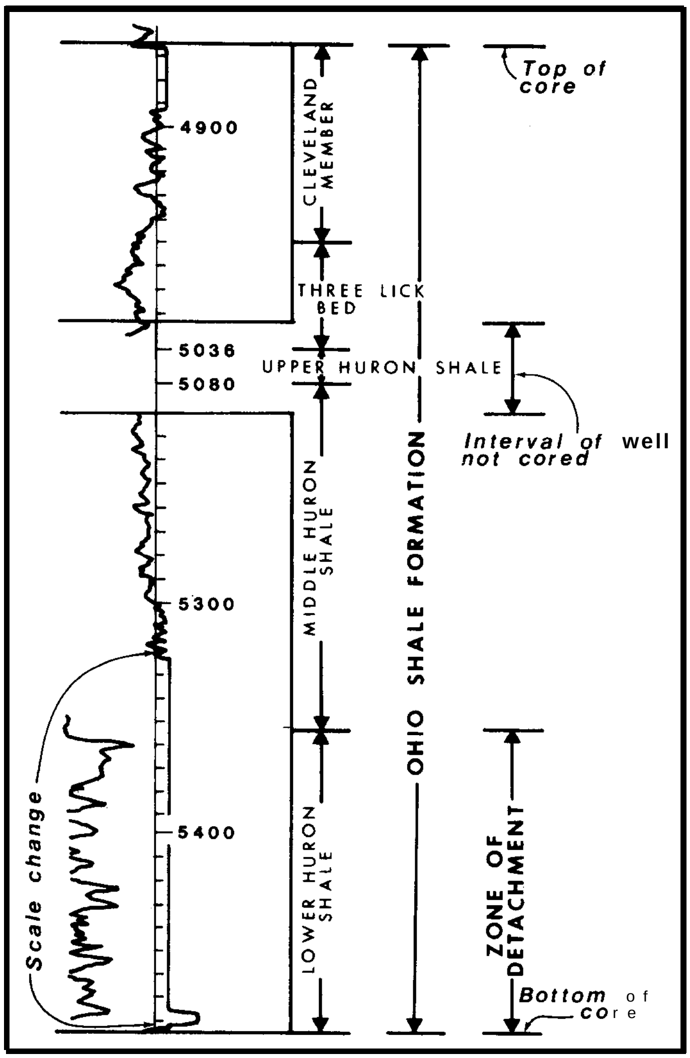


Figure 4 Equal area projections of poles to core induced fractures found in the Wise County well are shown for A) the most prevalent, unimodally striking core induced fractures (example taken from 4920 to 4930 feet of the core), and B) bimodally striking core induced fractures including the interval 5320 to 5370 feet of the core. Percentages of the total number of poles in the projections are indicated by tonal contrast.



GAMMA RAY LOG AND STRATIGRAPHY OF CORED INTERVALS WISE COUNTY, VA. WELL NO. 20338.

Figure 5

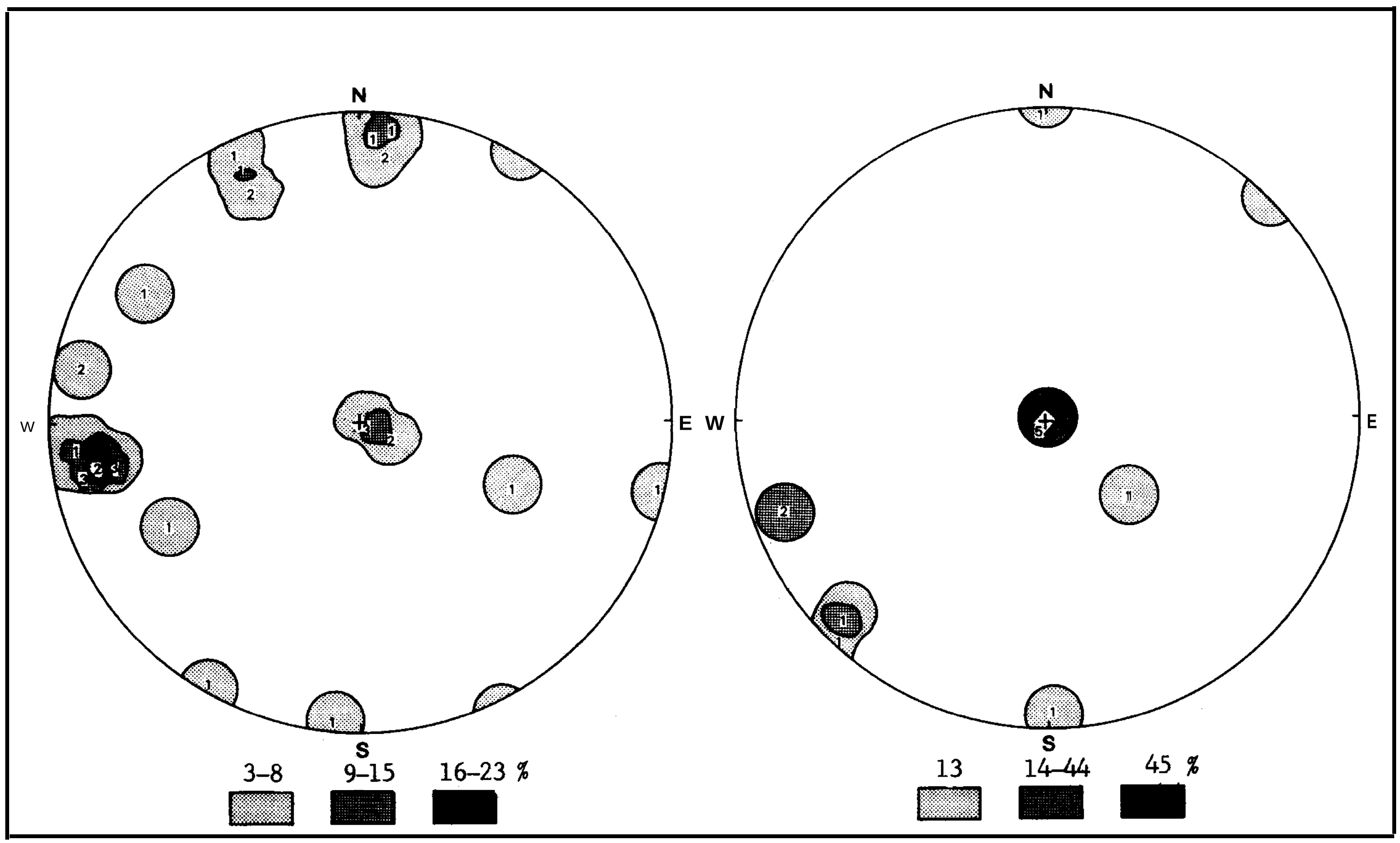


Figure 6 Equal area projections of A) poles to all natural fractures, and B) poles to all slickensides, found above the lower Huron shale in the Wise County well #20338. Numbers superimposed on projections are the numbers of identical orientations. Percentages of the total number of poles in the projections are indicated by tonal contrast.

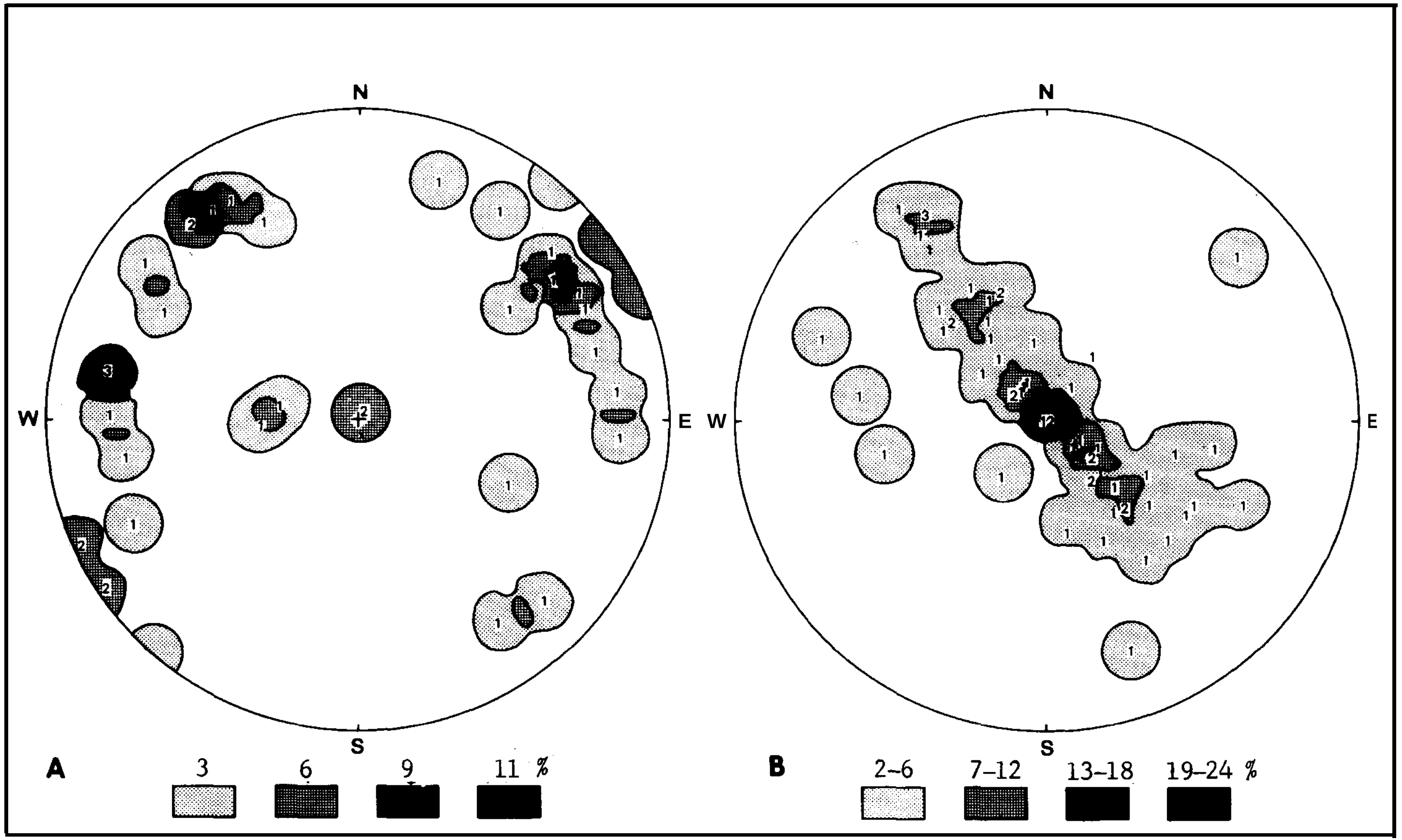


Figure 7 Equal area projections of A) poles to natural fractures, and B) poles to slickensides found in the lower Huron shale in the Wise County well #20338. Numbers superimposed on projections are the numbers of identical orientations. Percentages of the total number of poles in the projections are indicated by tonal contrast.

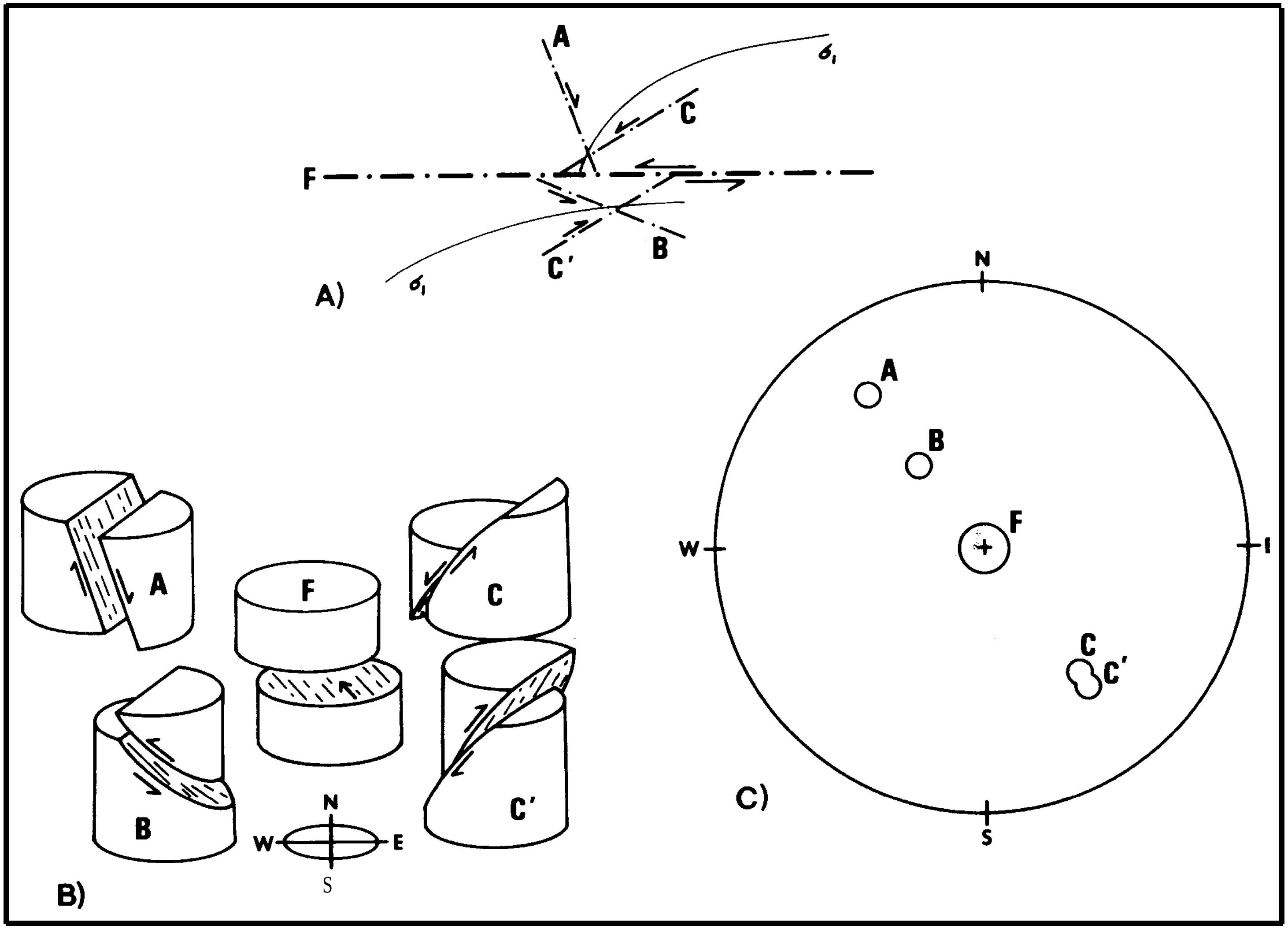


Figure 8 A) reoriented stress trajectories and the second order faults (A, B, C, C') predicted to form with major first order fault F (from Anderson, 1951). B) sections of core illustrating the orientations of the predicted first and second order faults for a NW vergent thrust F. C) equal area projections of the poles to the first and second order faults for a NW vergent thrust.

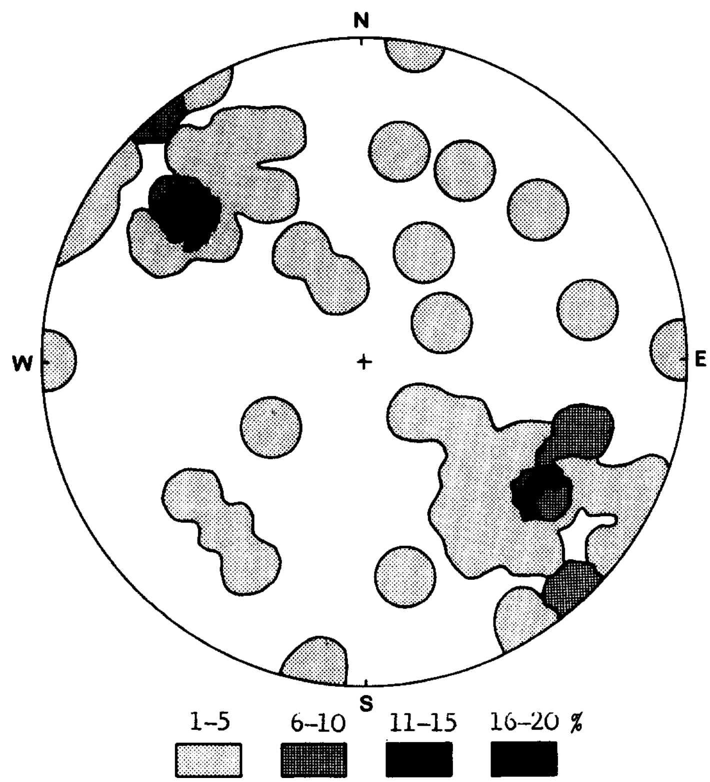


Figure 9 Equal area projection illustrates an anomalous amount of scatter in the poles to core induced fractures found from 2500 to 2550 feet in the Martin County (Kentucky) well #20336. Percentages of the total number of poles in the projection are indicated by tonal contrast.

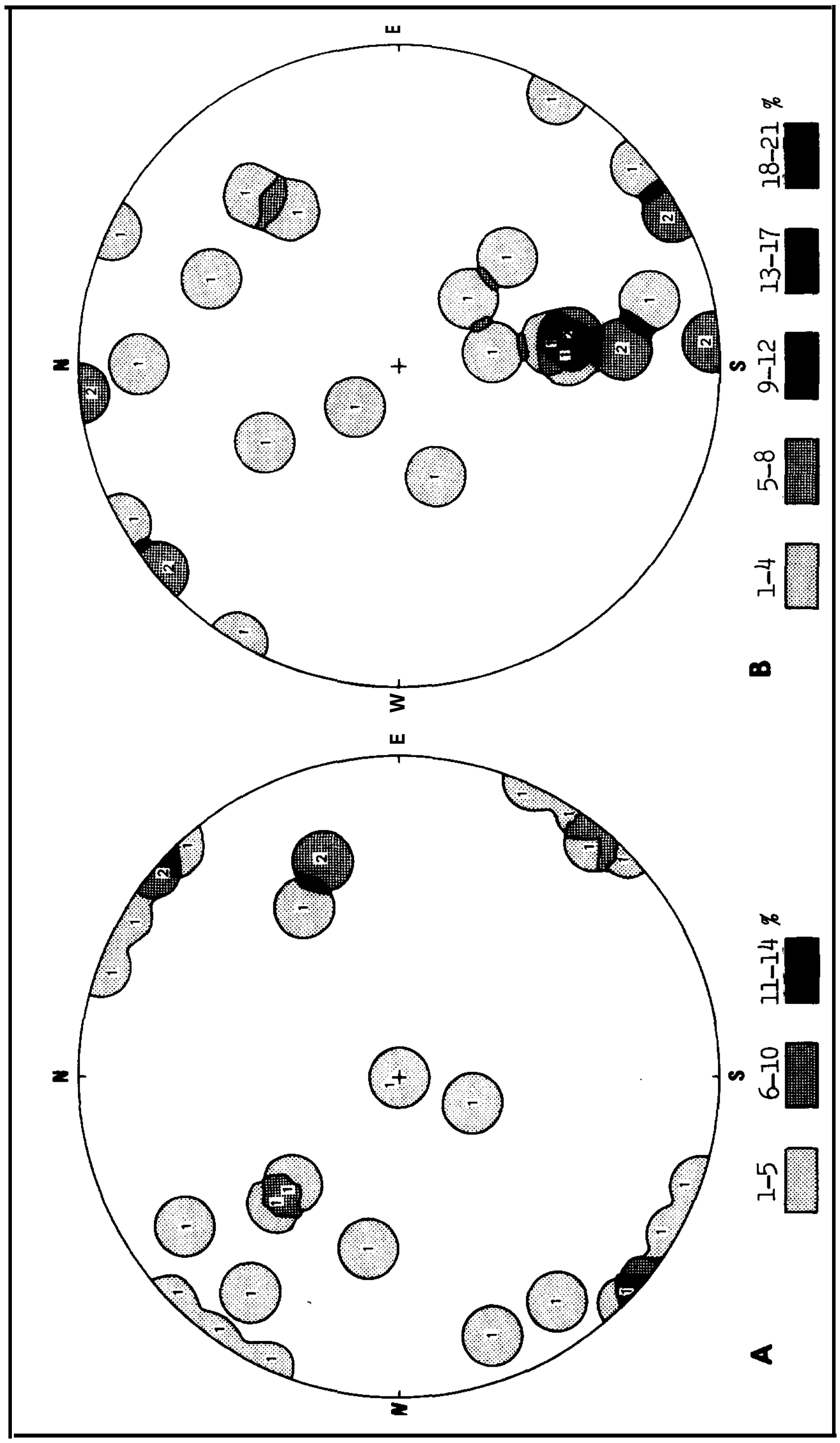


Figure 10 Equal area projections of A) poles to natural fractures, and B) measured and constructed slickenlines (see text) from the Martin County (Kentucky) well #20336. Numbers superimposed on projections are the numbers of identical orientations. Percentages of the total number of poles in the projections are indicated by tonal contrast.

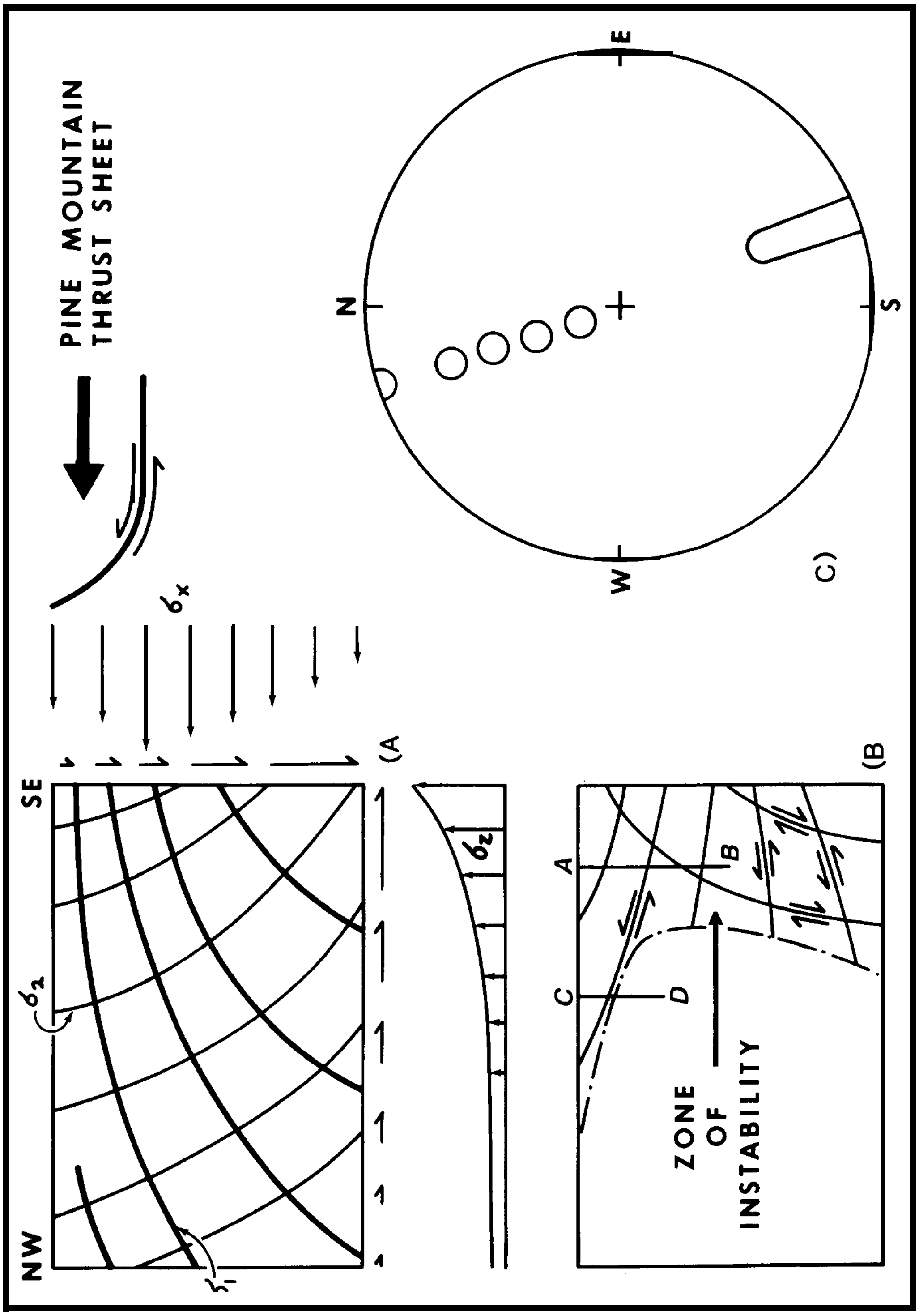


Figure 11 A) STRESS TRAJECTORIES PRODUCED BY EXPONENTIALLY DECREASING HORIZONTAL SUPPLEMENTARY STRESS (HAFNER, 1951). B) POTENTIAL FAULTS WITHIN UNSTABLE ZONE C) EQUAL AREA PROJECTION OF POTENTIAL SLICKENLINES.

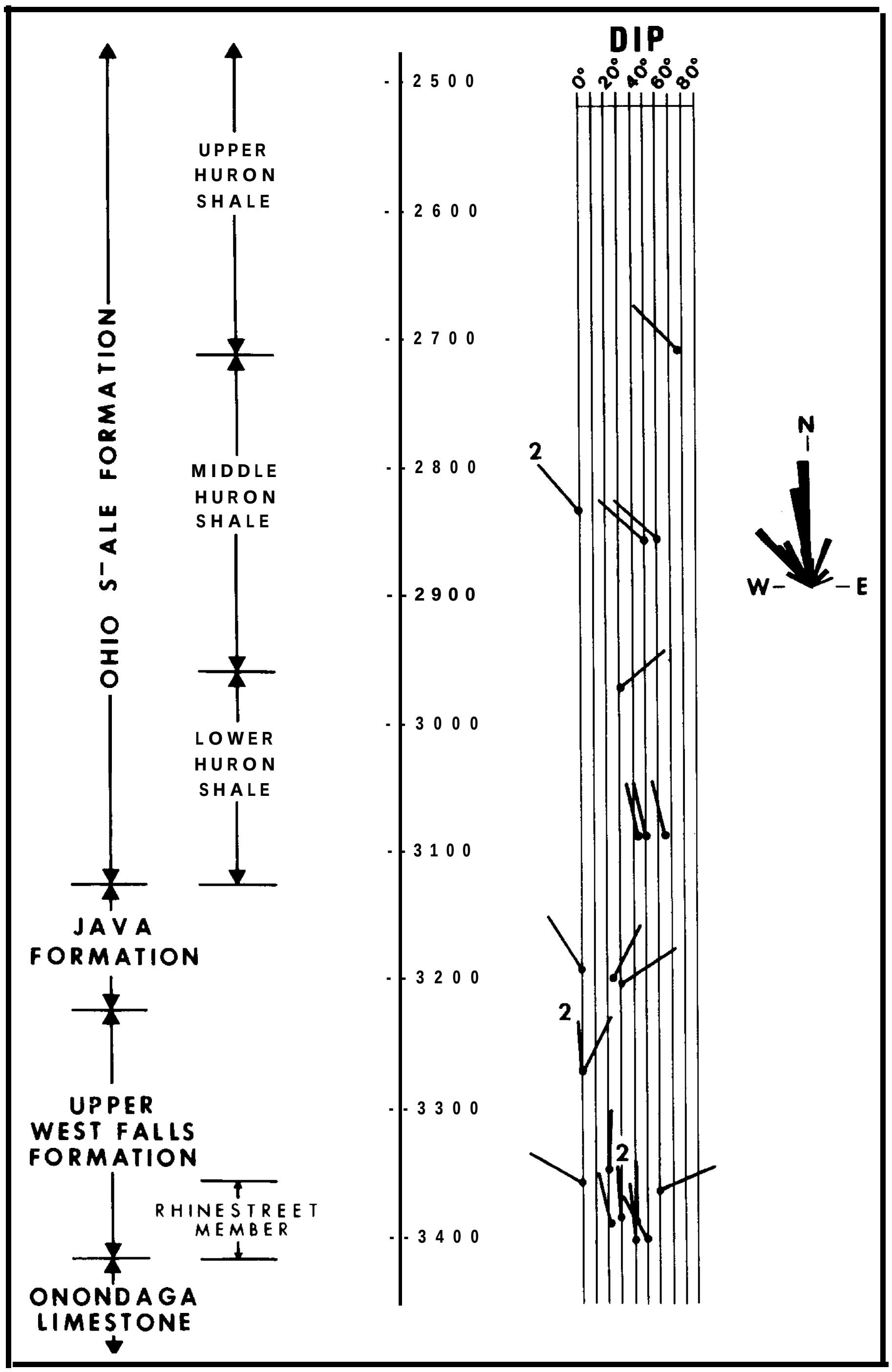


Figure 12

TREND AND PLUNGE OF SLICKENLINES, MARTIN COUNTY, KY. WELL NO. 20336.2 INDICATES TWO PARALLEL SLICKENLINES. ROSE DIAGRAM SUMMARIZES SLICKENLINE TRENDS.

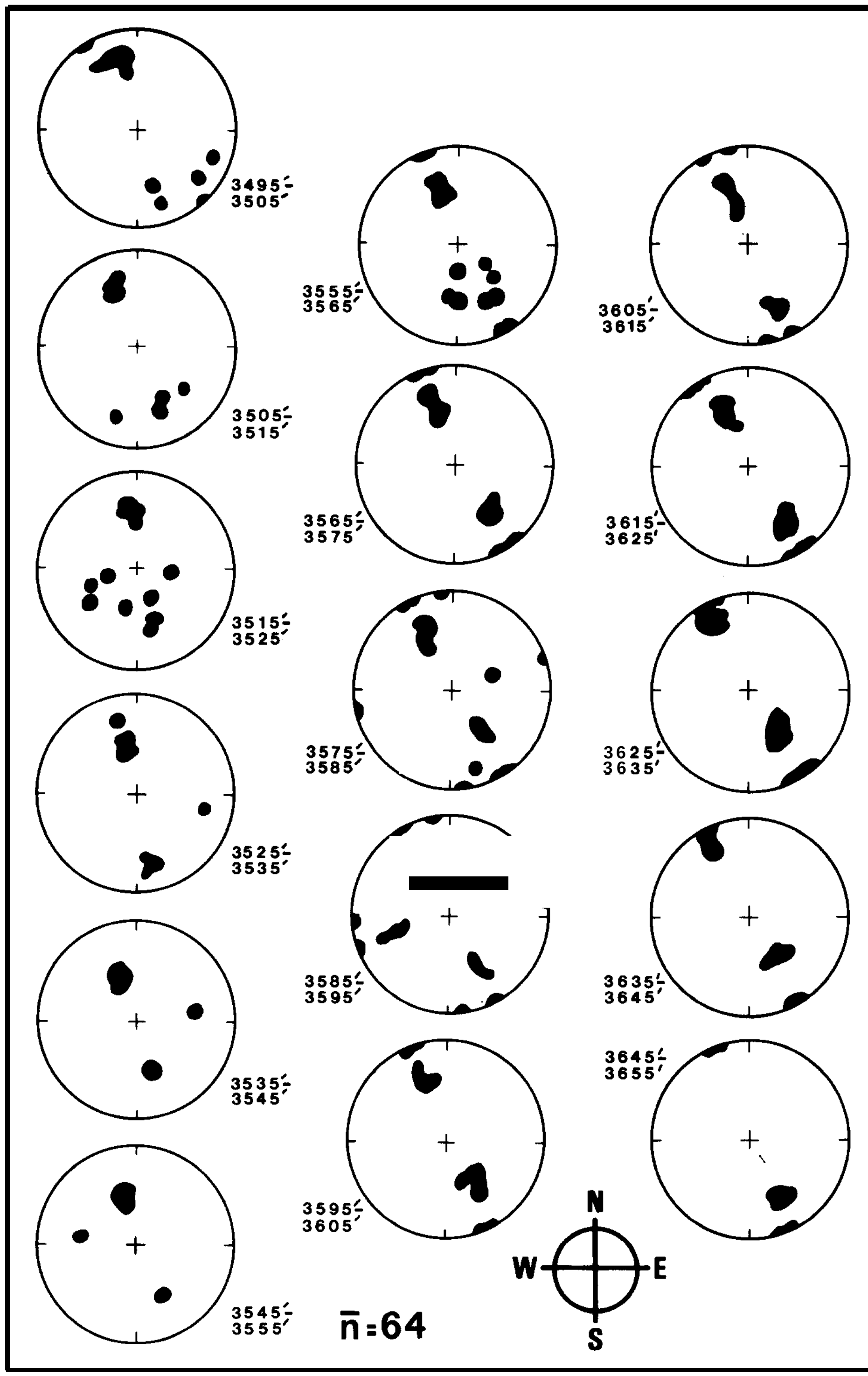


Figure 13

EQUAL AREA NETS LOWER HEMISPHERE: UNDIFFERENTIATED FRACTURES, MARIETTA, OHIO NO. R109 WELL.

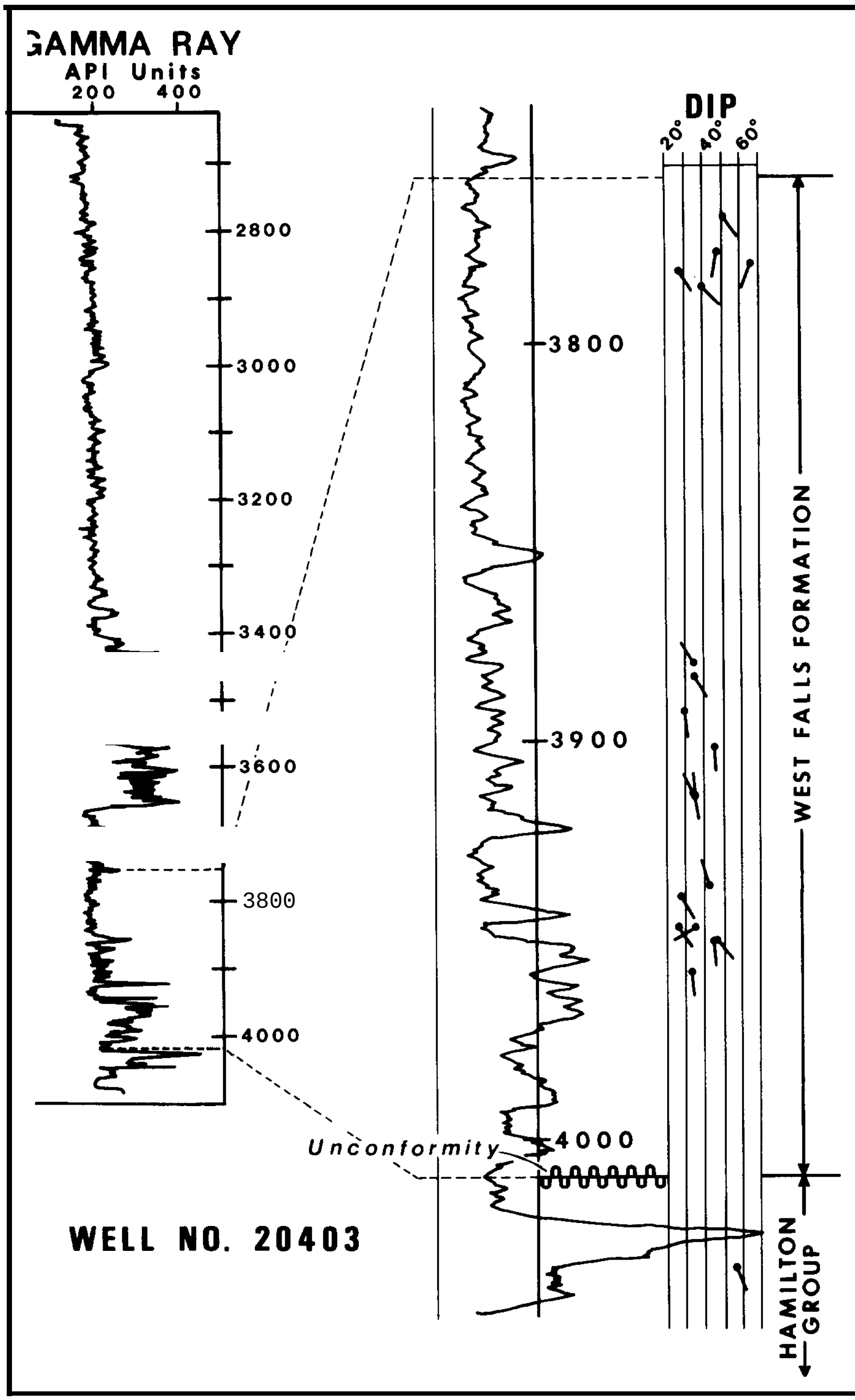
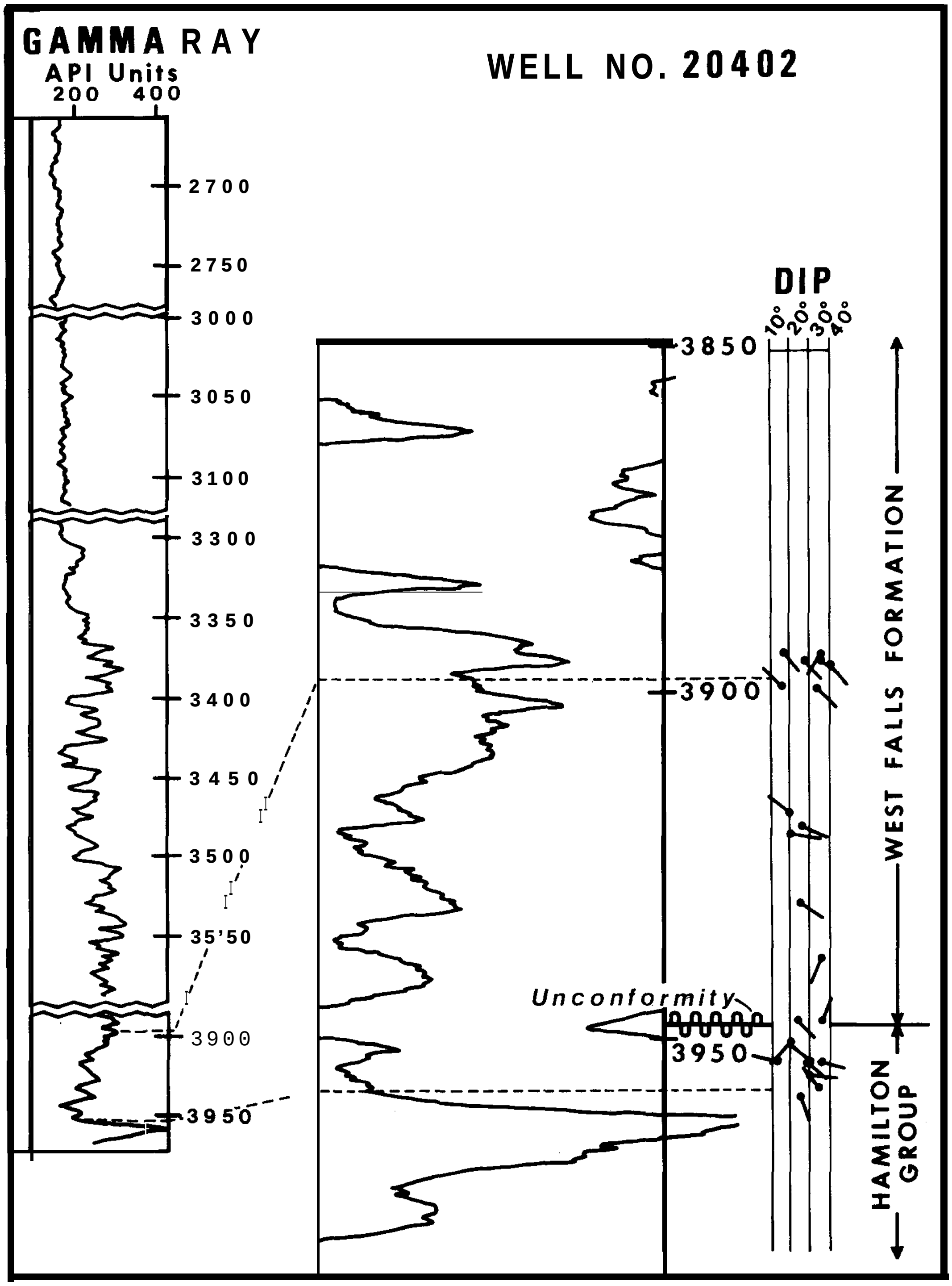


Figure 14

TREND AND PLUNGE OF SLICKENLINES, LINCOLN COUNTY, W.VA. ND. 20403 WELL.



.gure 15 Gamma ray log, stratigraphy, and dips of slickensided surfaces are shown for the Lincoln County well #20402. Cored intervals are within the West Falls Formation and the Hamilton Group.

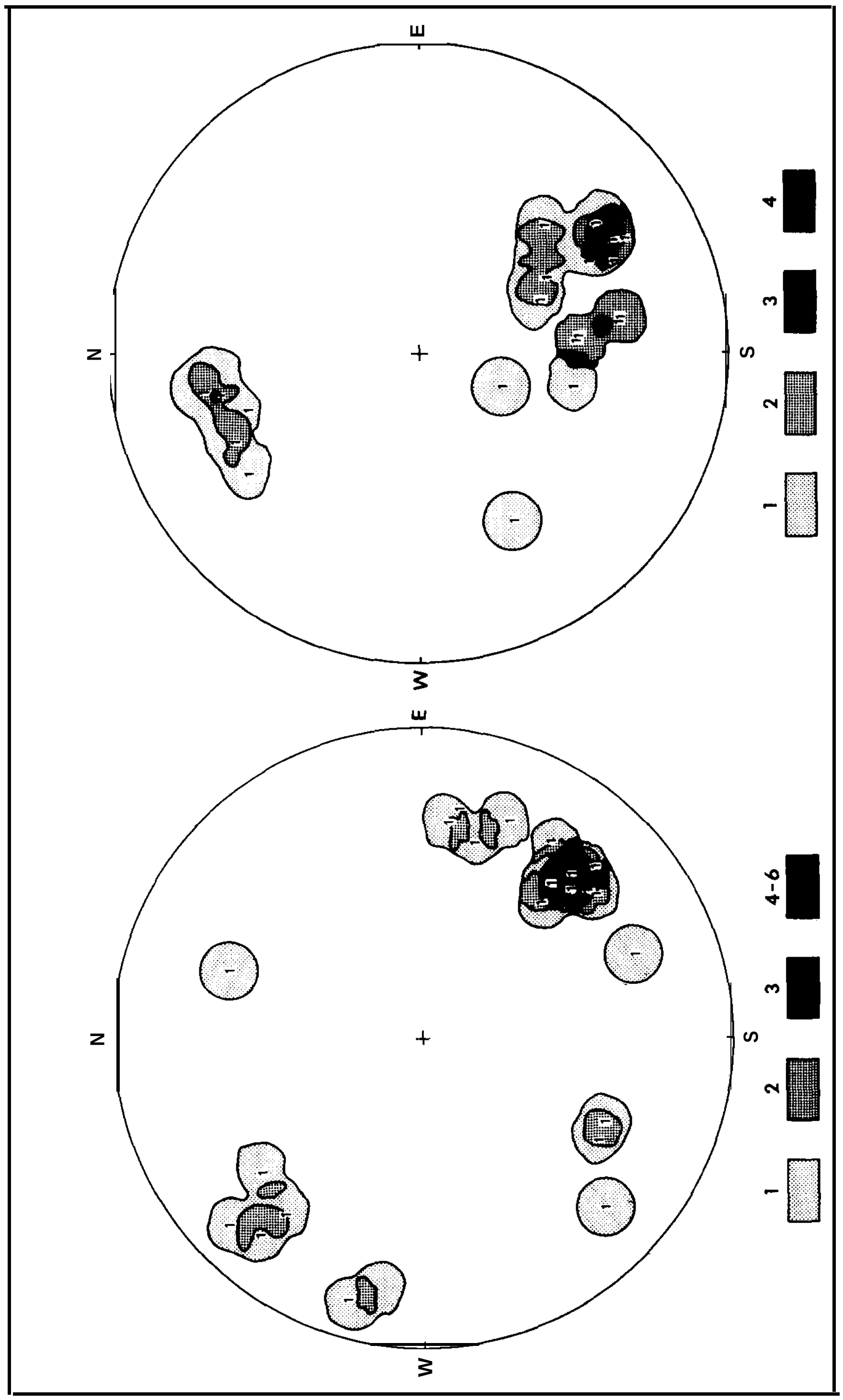


Figure 16 Mell is contour diagram of A) Lincoln County well #20402 and B) well #20403.
The numbers of poles are indicated by tonal contrast.

POROUS FRACTURE FACIES IN THE DEVONIAN SHALES OF EASTERN KENTUCKY AND WEST VIRGINIA

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ABSTRACT

Preliminary analysis of published fracture data obtained from seven oriented Devonian shale cores taken cooperatively by private industry and the U.S. Department of Energy, Eastern Gas Shales Project shows that fractures are numerous, uniquely oriented, and often mineralized in the highly organic portions of the Devonian shale of eastern Kentucky and West Virginia. These fractures and the commercial gas production generally occur together in the organic shales. We have applied the descriptive phase "porous fracture facies" to succinctly label the fractured and productive shale. Inclined slickensided fractures, found in the West Virginia cores, appear to be double shear sets that have an acute angle which opens toward the southeast. This orientation is approximately at right angles to the fold axes, which suggests that the fractures developed during the Alleghenian deformation are responses to differential shortening across the lower organic shales. A core taken in Kentucky contained horizontal slickensides in an upper organic shale. Those slickensides trend parallel to the thrust movement of the adjacent Pine Mountain thrust sheet. Inclined slickensided fractures trending North-South are found in a lower organic shale. Many problems remain, and more detailed analysis is being accomplished. The best of several working hypotheses contends that the fracture facies formed within a zone of abnormally high fluid (gas?) pressure located primarily in the lower organic shales. The western limit of abnormal high pressure may have been the basement fault zone along the western margin of the Rome Trough. The eastern limit of commercial production and open fractures appears to be the area of more intense tectonic transport in the area of the detached folds and thrusts.

INTRODUCTION

Through discussions with geologists interested in the nature of Devonian shale production, and from listening to papers dealing with fractures presented at the Department of Energy's First Eastern Gas Shales Symposium, one could easily gain the view that it is only vertically continuous fractures and surface fracture zones which are most important to the discovery of gas production from the shale. In my estimation, this impression is incorrect at least for the eastern Kentucky and southwestern West Virginia portion of Devonian shale gas production.

There is little doubt that fractures are important to production, but little has been said or written summarizing the present state of knowledge concerning the nature and origin of the fracture porosity responsible for production from the shale. Even though surface fracture zones, visible on imagery as lineaments, may be important in locating sites to be tested for enhanced production, we need to understand the type, orientation, and origin of the fractures responsible for gas production from the shale. If we do not document these characteristics, we will surely find it difficult to predict their extent and locations and, therefore, the extent and locations of production.

Fracture data from cores suggest that the free gas produced from the shale does not come from vertically continuous zones, but that it comes from fractures that have limited vertical extent found primarily in the lower highly organic portions of the shale section. Also these data, while admittedly few and not completely analyzed, show that the fractures are preferentially oriented and therefore stress related. Apparently the fractures are the reservoir, and the enclosing shale is both the source and seal for the gas entrapped in that fracture porosity. We have applied the descriptive term "porous fracture facies" to convey the image of a fracture porosity that has limited stratigraphic extent (de Wys and Shumaker, 1978a).

While the auther believes this interpretation is viable, he hopes that readers of this report will maintain a degree of skepticism. The data on which the interpretations are based are admittedly limited. Factors other than those structural which might influence production are not discussed. The importance of shale lithology, geochemistry, and fracture mineralization to shale gas production are affirmed. However, the author feels that the ideas presented above should have general circulation, for if correct, they will influence the direction of research within the project.

PURPOSE

The intent of this paper is to briefly summarize the state of knowledge related to the nature of fractures found in the productive zones. The published data pertinent to a generic interpretation for certain fractures is presented with the hope that other scientists within the project will be challenged to correct, to add to, or enrich the ideas presented. This is the first report concerning initial results of an on-going research project into fractures in the shale.

Stratigraphic Distribution of Production

By now, most scientists working with the productive Devonian shale generally accept the view that gas production primarily comes from fractures and the organic portions of the shale. Several papers present data in support of this conclusion.

Perhaps the first paper dealing with this subject was written by Hunter and Young (1953) who discussed the relationship of natural gas production to joints and fractures in the Devonian shale of eastern Kentucky. Bay (1968) clearly showed the association of gas production with zones of logged high radioactivity in the shales of eastern Kentucky. Martin and Nuckols (1976), and Patchen and Larese (1976) not only demonstrate the applicability of these observations to the productive shale in the Gottageville (Mount Alto) field of western West Virginia, but they also discuss the importance of mineralized vertical fractures found in the lower organic shale to gas production. This association of organic layers, fractures, and production can be demonstrated by the comparison of descriptions of cores taken in southern West Virginia through analysis of several reports by Byrer et al. (1976) (Figures 1 and 2). Larese and Heald (1977) discuss the petrography of the shale in two cores from West Virginia, and they emphasize the importance of vertical permeability associated with mineralized vertical fractures and horizontal permeability associated with silt beds within the shale. They also show one micro shear fracture which presumably had little porosity associated with it. Based on these data, it seems inescapable to conclude that present commercial gas wells produce primarily from the fractures very often found in the organic-rich portions of the Devonian shale. This conclusion does not preclude production coming from open fractures at other stratigraphic levels (see Figure 2), nor does it contradict the view that vertical zones of intense fracturing visible as lineaments, could influence shale gas productivity. However, the occurrence of stratigraphically limited fractures does make one seriously question any inferred identity between the orientation, character, or even the origin of surface fractures with those found in the productive zone. Indeed, Werner (1977) presented preliminary data which suggest a suspected inverse relationship between Landsat lineaments and the shallow shale gas productivity for southern West Virginia.

Generally, the production from fractures of the organic zones has been correlated with increased fracture density and mineralized fractures in those stratigraphic intervals. However, little has been said of the evidence from the cores that there are inclined slickensided fractures in and around the lower organic shales, and even less has been said about their origin. These three changes in fracture characteristics, that is, changes in density, mineralization, and orientation, form the basis for the proposition that a porous fracture fhcies is present in the lower organic shales, and that an understanding of fractures is important for understanding production from the shale.

Fracture Facies: An Interpretation

Thus far I have briefly summarized what is known of shale fractures obtained primarily from descriptions of the first few cores collected during the initial stages of the Department of Energy's Eastern Gas Shales Program. The data points are few, and they are concentrated in commercially productive areas of West Virginia and eastern Kentucky. Therefore a broadly applied interpretation based on these few data is hazardous. Nonetheless, I believe that the data are sufficient to suggest a preliminary interpretation for the origin of certain of the fractures found in the organic shales of eastern Kentucy and southern West Virginia. Furthermore, I believe that it is worthwhile to propose this tentative interpretation because in the process of assessing the hypothesis one is challenged to think critically about collection and analysis of fracture data. In addition, a working hypothesis gives one a standard against which ideas can be tested and integrated with other scientific interpretations concerning the nature of the shale production and shale structures. Finally, and perhaps most importantly, a generic hypothesis such as the one proposed here lets one predict the total character and extent of fracture porosity, and it helps one predict the effect which a fracture facies might have on experiments and production tests run by other scientists working in the Eastern Gas Shales Program. I will point out a few such ramifications that I can foresee later in the text.

Shumaker (1976) proposed that at least part and possibly all of the fractures responsible for shale gas production in eastern Kentucky and southwestern West Virginia were caused by minor tectonic transport and/or differential shortening within and above an ancient high-pressure detachment zone in the basal Devonian shales of the "undeformed" Appalachian foreland.

High pressure under-compacted shale intervals, similar to that proposed, serve as conduits for dewatering of sediments. Such zones often form basal glide surfaces for the gravity tectonic structures found in areas such as the Gulf Coast of the United States. Hubbert and Rubey (1959) proposed that the high pressure fluid reduces friction in the containing layer to permit the movement of overlying rock for great horizontal distances during thrusting. The Hubbert and Rubey hypothesis, as it is often called, is clearly applicable to movement and deformation in the Pine Mountain thrust block (Figure 1), and it is presumably applicable to the detached folds found directly east of the shale production (Figure 1) in West Virginia. It is also reasonable to presume that the stress field and high pressure zones at the base of the shale associated with the deformation extended westward into the undeformed foreland across the shale gas area. The stress associated with minor differential shortening across the basal shale could have created certain of the fractures of the porous fracture facies. What we know thus far about characteristics of fractures even to the mineralization, suggesting abundant water movement, is compatible with the proposed interpretation.

Evidence in Favor of Differential Shortening

The characteristics described thus far could also be explained by most other theories for the origin of fractures (Hodgson, 1976). For instance, it is the rule that bed lithology affects joint spacing and sometimes even the presence of mineralization. One might expect under normal conditions that the organic portions of the shale would have their own characteristic joint spacing and mineralization as compared with the more silty portions of the shale section. One should ask then, what is unique in the characteristics of these fractures which support my contention that at least part of the fractures of the porous fracture facies were created by differential tectonic transport or differential shortening across the basal shales? I have listed below the data and observations which I believe support this interpretation above other possible alternatives:

1. Slickensides indicative of tectonic transport are reported from the shale, particularly in the lower organic zone. The presence of low-dipping slickensided fractures in the Lincoln County cores of West Virginia (Figure 1) were analyzed by J. Dixon who found inclined fractures (personal communication)[2] which probably are complimentary shear fractures. At least one fracture system is present. An approximate 60° angle between the two shear sets of the system opens toward the southeast, approximately at right angles to the fold axes. Therefore, I feel that this system formed in response to the same stress field as that which formed the adjacent folds. Additional work will be required to explain the entire fracture pattern found in the cores and to assess its relationship to the surface joints. However, it appears that horizontal slickensides are more numerous near the Pine Mountain thrust, and that the shear fracture system predominates in the distal areas of movement.

2. The proximity of gas production to known detachment within the basal organic shale suggests an association. The Pine Mountain thrust block, detached within the same basal shale, lies directly east and south of commercial production. However, the areas of intense transport such as the Pine Mountain thrust and the more folded areas of West Virginia have not produced significant amounts of commercial shale gas. The present eastern limit of commercial production approximates the margin of more extensive tectonic transport. The western limit of commercial shale gas production approximates the northwestern margin of the deeply buried Rome trough (Figure 1). Although Cambrian in age, minor reactivation of the trough has had continued effect on sedimentation and surface structures visible in Pennsylvanian rocks. It could be interpreted that these structural boundaries mark the limit of the most intense development of the porous fracture facies caused by shortening within the shale. It has been suggested (B. Long, personal communciation) that the vertical fault might have provided avenues for release of water pressure from the high pressure zone, thus indirectly providing a western limit to the fratture facies. However, this western limit of commercial production may also reflect other limitlng geologic-geochemical factors such as the distribution of thick organic-rich shales. Definitive conclusions clearly are premature, and only empirical relationships are noted here. Indeed, these empirical relationships may simply reflect economic factors such as rate of return on wells drilled, but clearly the best wells fall within these structural boundaries.

3. Kulander et al. (1977) also reported horizontal and inclined slickensides from the Perry County well of Kentucky (Figure 1). The slickensided horizontal fractures are reported to be evenly distributed throughout the core interval. There is a variety of slickenside trends logged within the core by Kulander et al. (1977), but one of the prominent trends for horizontal slickensides is in the same northwest-southeast azimuth that is compatible with the inferred northwestward tectonic transport of the Pine Mountain thrust block. Other prominent directions of movement include north-south trending inclined fractures found in the productive lower organic shale zones and three horizontal slickensides aligned in a northeast-southwest direction. In undeformed shales, slickensides are often called compaction features; however, if the slickensides were formed by compaction, then they should be random in trend.

4. The sequence of fractures found in the Wise County, Virginia, shale core taken through the upper plate of the Pine Mountain thrust block is similar to that seen in the cores taken in "undeformed" foreland productive zone in that the fractures are inclined and horizontally slickensided surfaces that are mineralized (Wilson, et al., 1978). The indications of movement and fracturing are far more intense in this core where major tectonic transport (a minimum of eight miles) along the bedding fault is clearly established. Most significantly, the detachment appears to be in the lower organic shales. When the detachment zone is encountered by the drill there usually is a gas blow indicating abnormal pressure (Young, 1957); but unfortunately the gas appears trapped in isolated pockets to produce only non-commercial quantities.

5. Finally, there is the widespread character of shale gas production in Eastern Kentucky and extreme southwestern West Virginia which seems to suggest some horizontal transmissability of the gas throughout the productive area. While there is great variety in production from well to well, the chance is slight that each well drilled in the area of production would intercept a vertical fracture. Published descriptions of all cores from productive wells in West Virginia and eastern Kentucky include a notation of slickensides. Most of these well descriptions and photographs show inclined fractures. Presumably these fractures would impart some horizontal permeability; however, it is not yet certain if inclined fractures are as significant or if they are more significant than other factors such as silt layers (Larese and Heald, 1977).

Ramifications of the Differential Shortening - Fracture Facies Hypothesis

The hypothesis proposed above for the origin of certain fractures of the porous fracture facies should affect shale production characteristics research or experiments associated with the Eastern Gas Shales Project in the following ways:

1. One would expect the presence of mylonite and tectonic breccia within the basal organic shale section. Thus far, breccia has not been described from this zone, but thin fine mylonite is occasionally noted on fractures with slickenside surfaces from descriptions of the cores.

2. High pressure gas zones within the basal shale section should be present. Some high pressure blow-outs are reported, but no systematic study of the distribution has been undertaken. In the Pine Mountain thrust block the presence of high-pressure gas pockets encountered by the drill has been noted for years.

3. Fracture intensity should decrease northwestward away from a maximum zone of intense deformation. This trend should approximate the trend of the adjacent Appalachian structure or basement structure that could influence stress trajectories. Data are not **yet** available on this. It is not. yet clear why the shale is essentially non-productive eastward in the areas of extensive tectonic transport. **The** western margin of commercial production appears to be close to the western margin of the Rome trough. The significance of this is not clear, since the trough affected sedimentation and thus production could reflect either stratigraphic or basement structural control.

4. The blanket-like characteristic implied with the fracture facies hypothesis should produce traps similar to stratigraphic traps near its margin. Within the central area of maximum development, structurally high areas should be generally more productive than low areas.

5. Horizontal tranmissibility of gas either through fractures or other means should be reflected in production characteristics. Wells show some degree of pressure interference that is not strictly linear in trend as would be expected with simple vertical fracture interconnection.

6. Where fractures, created by other stress fields, intersect the fracture facies one should obtain either abnormally high or low producing wells if a seal or breach occurs above the intersection of the fractures. If the zone of intersecting fractures is sealed from the surface, then abnormally high production should occur along that trend. If the vertical fractures are not sealed, then gas should escape to the surface and the productive trend should be abnormally low.

7. Any of the usual causes for creation of more intense fracturing would enhance porosity in the fracture f **acies.** Basement flexures or pre-existing structural features might create abnormal edge conditions within the stress field at the base of the shale to cause enhanced fracture density and thus abnormally productive wells. Such areas might also give anomalous, deviatoric stress conditions that would produce reoriented residual stresses within the shales to cause uniquely oriented fractures (Overbey, 1976; Advani et al., 1978).

8. Deviated wells should not be designed solely on the presumption of encountering vertical fractures alone. The design should be tailored to the structure of the area, presumably only after an oriented core is taken in the area to define the fracture characteristics. It is imperative that the fractured portion of the shale (e.g., the lower organic shale) be defined for intersection by the inclined bore hole. This restricts the target to be intersected within the shale to one of several tens of feet vertically.

9. Any induced fracture program must be designed to take into account reoriented stresses and different types of fractures in the different lithostructural units of the shale section.

10. It seems unlikely that the stress fields related to the formation of the inclined slickensided fractures of the facies extended westward beyond the limit of basement arches that outline the Appalachian Basin; and as mentioned above, it may have extended only to the western edge of the Rome trough. If certain shale beds were over pressured across the continental interior, then other more local stress could produce the local development of a fracture facies. **The** Cottageville and Midway-Extra fields of West Virginia may be the best examples where such local intensification of fracturing has occurred over basement structures. Clearly this model or exploration rationale is usable within all basins, but it is most applicable to the interior basins e.g. Illinois, western Kentucky, Michigan, etc. Such basins should not have large areas of horizontal transport and shortening across the Devonian Shales.

11. Very detailed structure maps are necessary to define productive structures of the types described in (10) above. For instance, a 25 foot contour interval may not isolate these structures from regional dip, but a lo-foot interval may. Of course, elimination of regional dip and trend surface analysis may assist in this definition.

12. At present, we cannot predict precisely where a porous fracture zone (porous facies) will occur as we have encountered both high and low side structure-productive trends. Certainly models such as that described by Advani et al. (1978) will be helpful, but study of actual productive analogs is essential (see de Wys, Nuckols, this Proceedings and Schaefer thesis in progress) [3] if we are to improve on these exploration rationales for the shales.

CONCLUSION

The descriptions of core data, primarily accomplished by Department of Energy and West Virginia Geological Survey personnel, and the work of several investigators within the Department of Geology at West Virginia University regarding fractures within the shales support the view that there are more and uniquely oriented fractures and more abundant fractures within the basal shale section when compared with the remaining shale section. This porous fracture facies, as it is called, is generally found in or around the lower organic shales, and it is considered to be critical for commercial production found in eastern Kentucky and southern West Virginia. The limited data generally support the hypothesis that at least the horizontal and inclined fractures of the facies developed as a result of differential tectonic transport or differential shortening across an ancient high-pressure zone found in the organic shales. Understanding the origin of open fractures and the fracture facies is leading toward a rationale for exploration that is based on productive fields, and we may be leaving the random drilling phase of exploration and entering a period where exploration is based on successful analogs that can be applied to non-productive but geologically and geochemically appealing areas. Eventually we may be able to explore for and directly detect open fracture zones.

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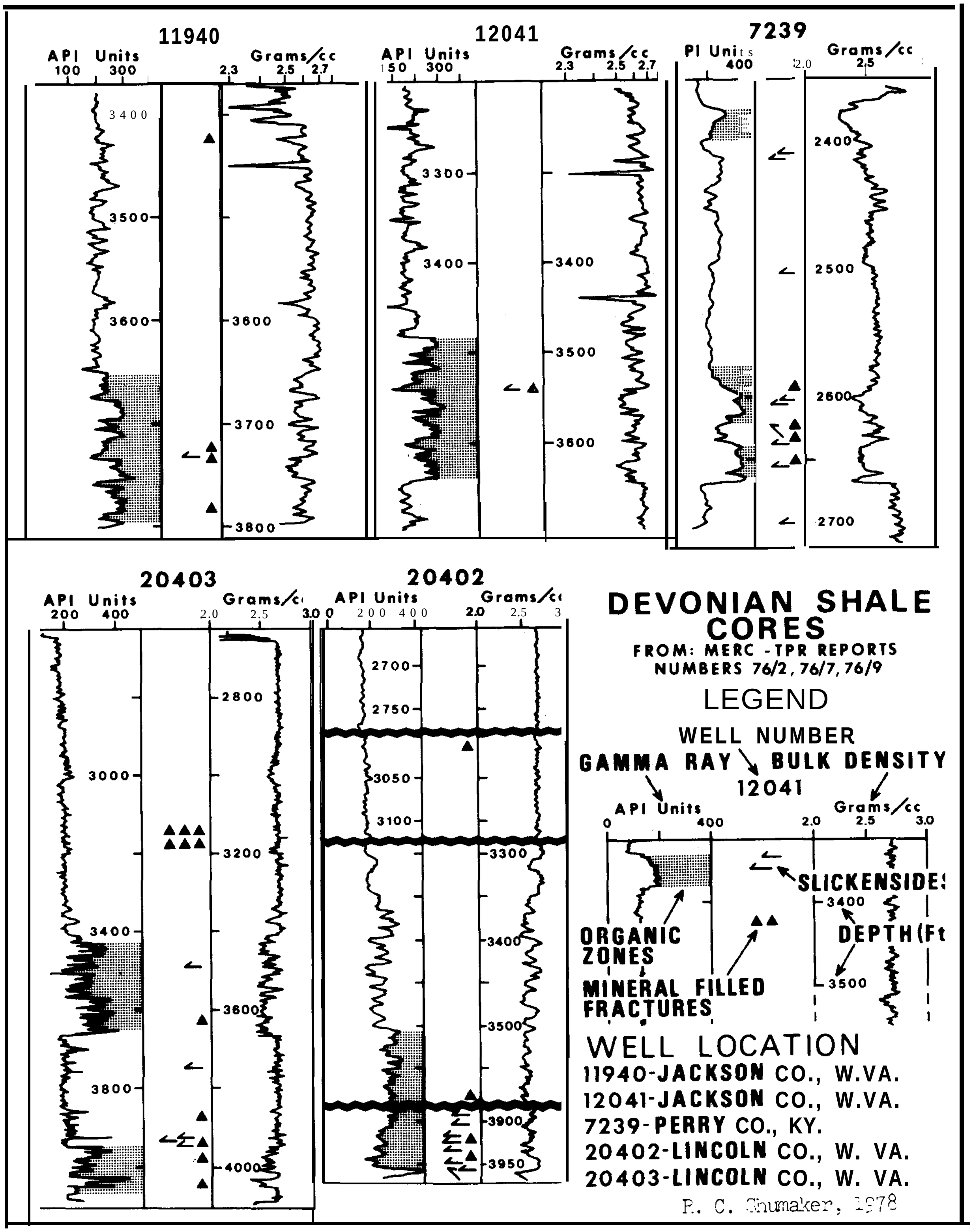
11940
API Units
Grams/cc
12041
7239
20403
20402
DEVONIAN SHALE CORES
FROM: MERC -TPR REPORTS
NUMBERS 76/2, 76/7, 76/9
LEGEND
WELL NUMBER
GAMMA RAY
BULK DENSITY
12041
SLICKENSIDES
DEPTH (Ft
ORGANIC ZONES
MINERAL FILLED FRACTURES
WELL LOCATION
11940-JACKSON CO., W.VA.
12041-JACKSON CO., W.VA.
7239-PERRY CO., KY.
20402-LINCOLN CO., W. VA.
20403-LINCOLN CO., W. VA.
R. C. Shumaker, 1978

Figure 1

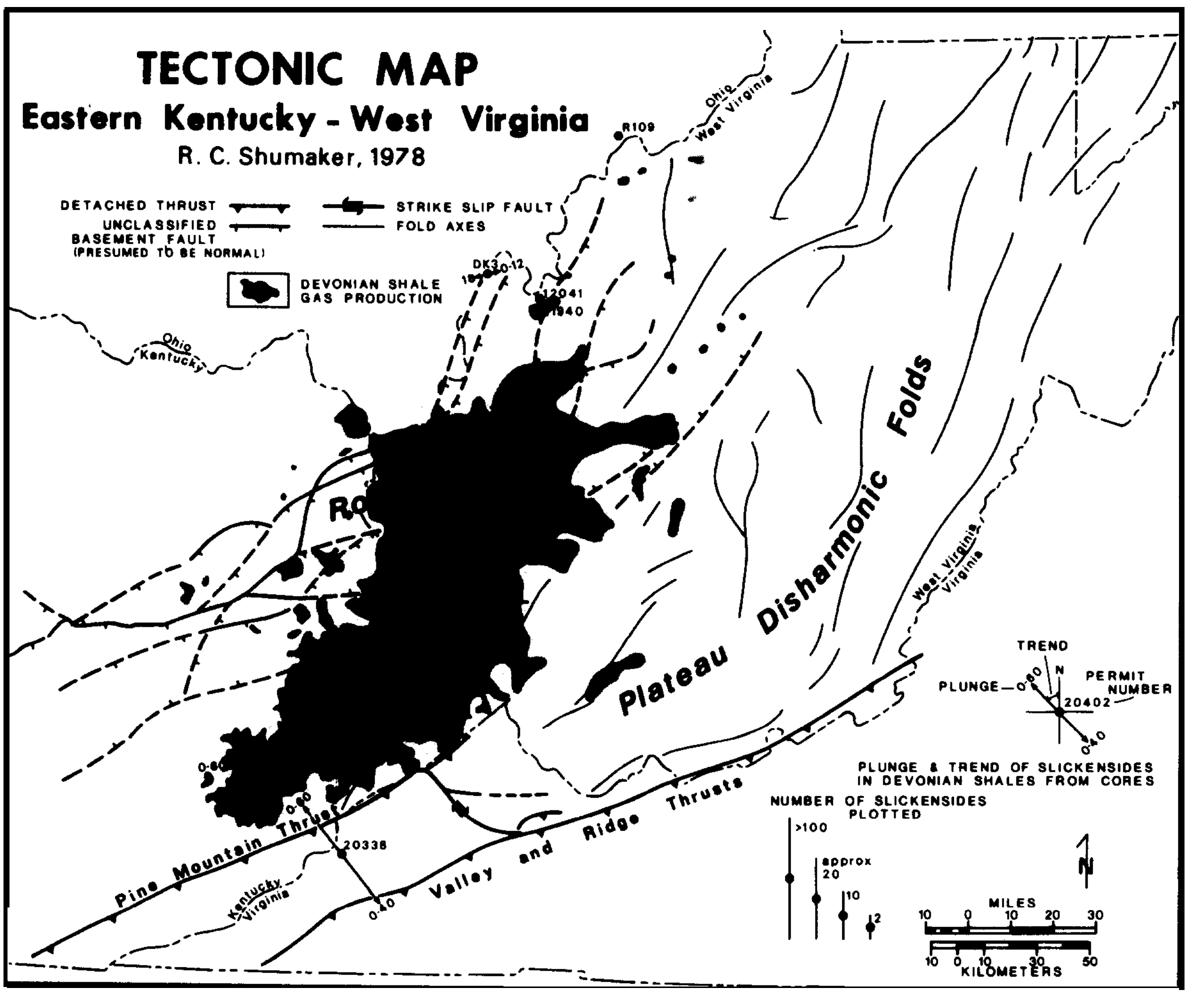
TECTONIC MAP
Eastern Kentucky - West Virginia
R. C. Shumaker, 1978
DETACHED THRUST
UNCLASSIFIED BASEMENT FAULT (PRESUMED TO BE NORMAL)
STRIKE SLIP FAULT
FOLD AXES
DEVONIAN SHALE GAS PRODUCTION
Ohio
West Virginia
Ohio
Kentucky
R109
12041
Plateau Disharmonic Folds
West Virginia
Virginia
TREND
PLUNGE
PERMIT NUMBER
20402
PLUNGE & TREND OF SLICKENSIDES IN DEVONIAN SHALES FROM CORES
NUMBER OF SLICKENSIDES PLOTTED
>100
approx 20
10
2
MILES
KILOMETERS
Pine Mountain Thrust
Valley and Ridge Thrusts
Kentucky
Virginia
20338

FIGURE 2

BRITTLE FRACTURE PATTERNS ON A TRAVERSE OF THE BIG SANDY GAS FIELD, KENTUCKY

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ABSTRACT

The Big Sandy gas field of eastern Kentucky represents fracture production from Devonian shales. A series of remarkable new road cuts traverses the field along the Big Sandy River to provide the opportunity to correlate various classes and scales of brittle fracture elements across it. Common joints and coal joints occur in orientation domains semi-independent of each other but related in part to the irregular hinges of the West Virginia system of folds dying southwestward into the area. Most regional topographic lineaments appear to be systems of brittle fracture elements superimposed on a much broader area, possibly as subtle extensile features associated with subcontinental stress trajectories. Possible field evidence for only two of the lineaments was detected, both being joints associated with very small scale thrust faults. Joint zones or "hill seams" in one huge cut appear to be surface geomorphic effects of the upper 100 meters associated with gravitative collapse and lateral spreading of mechanical units bounded by coal strata. The model may have some practical application in coal mining safety and highway engineering of the area. There is a surprising lack of indication of Pine Mtn. thrusting effects in the area, even to the near absence of small scale slickenside surfaces. Instead the brittle fracture elements of this part of the gas field, like the structural contour pattern, should be considered the dying end of the West Virginia folds rather than the foreland of the Pine Mtn. thrust.

INTRODUCTION

The Big Sandy gas field of eastern Kentucky owes much of its production to fracture permeability in Devonian shales. The field is located among a number of structural trends of the Appalachians (Figure 1): the Pine Mountain Thrust at N62E, the West Virginia segment of the Appalachian folds at N40E, and the Kentucky River and related fault zones at N80E. Much of the field is within the Jenkins 2° Quadrangle.

Recent improvements on highway route 460 through Pikeville and Prestonburg, Kentucky, (Figure 1) have produced phenomenal road cuts on a line across the field. Cuts comprise about 25% of the distance along the line with individual cuts commonly 50 to 75 meters deep and up to a km in length. The largest cut involves production of a man-made "canyon" cutting off a meander at Pikeville. The cut is a km in length and up to 165 m (550 feet) deep. The cuts are largely in the Pennsylvanian Breathitt sandstones, conglomerates, and coal measures. Most are located on published U.S.G.S. 7 1/2' quadrangle maps which include structural contour data. These U.S.G.S. maps have been combined into the structural contour map of Figure 2, illustrating the structures between the Pine Mtn. thrust and the Irvine-Paint Creek fault of the Kentucky River system. The West Virginia fold trends end southwestward across the map in a complex NW-trending zone. This zone in the vicinity of the Prestonsburg Quadrangle is along a line generally followed by the Big Sandy River. The largest fault found in this investigation is a NW-trending thrust located approximately in the Prestonsburg zone just north of the Pikeville Quadrangle (Figure 2) on the Pike County-Floyd County line. The bedding plane nature of the fault, its two meters of telescoping, and its anomalous strike for this region are illustrated in Figure 3.

This paper is a progress report on brittle fracture investigations along the Pikeville-Big Sandy line based on field work in the summer of 1977. Field work was supported by a research grant from Mobil Oil Co.

TOPOGRAPHIC LINEAMENTS

The deeply etched plateau country of eastern Kentucky reveals many topographic lineaments in most forms of regional imagery. A wind rose plot of 415 lineament orientations drawn from a Band 7, Feb. 12, 1976, LANDSAT image of the Jenkins Quad is illustrated in Figure 4. The absence of NW-trending lineaments is interpreted as a low angle lighting effect produced by the lack of shadow enhancement parallel to the sun azimuth of 141 degrees. The reality and the accuracy of orientation of the NE-trending lineaments (which average only 10 km in length) are also suspect because of shadow illusions in a single low angle lighting direction.

The topographic lineament pattern was also investigated using shadowed relief maps as described by Wise (1969). Lineaments were drawn for relief maps of the Jenkins and nearby quadrangles as shadowed in four different lighting directions. Lineaments appearing in several lighting directions (Figure 5) are probably real and appear to be parts of systems traceable from 2" quadrangle to 2° quadrangle. Cumulative plots for all four lighting directions for the quadrangles are presented in wind rose form in Figure 6 with cumulative number in the upper half and cumulative length in the lower half of each rose. Many of the lineament trends are correlatable from rose to rose. Some may be related to local fold trends as, for example, the sets approximately parallel and normal to the Pine Mtn. thrust. The relations are permissive, not conclusive. Most of the lineament patterns seem to represent domains of far broader extent than this segment of the Appalachians. It will require coverage of a much more extensive region to examine what relationship, if any, exists between these lineament trend domains and the more traditional structures and structural domains of the Appalachians. We regard most of them as subtle extensile effects associated with stress trajectory traces of subcontinental dimensions (Wise, 1967).

A search for the physical manifestation of these lineaments was made in the huge road cuts with only two possible correlations: (1) The Big Sandy River flows along a diffuse and poorly defined NW-trending lineament. The thrust fault, illustrated in Figure 3, is subparallel to the lineament, as is the break in structural contour patterns in Figure 2. (2) A better correlation exists in the Pikeville "canyon" cut illustrated in Figure 7. The "canyon" transects Poor Farm Hollow, a linear tributary valley of the Big Sandy of approximately one km length with a strike of N20E. In the deepest level of the "canyon" a SE-dipping thrust fault of similar strike splays upward out of bedding to disintegrates into a series of irregular joints oriented N20-25E. These joints are well-developed, occur far below the depth of most jointing in the cut, and are immediately beneath the line of Poor Farm Hollow. For so major a topographic effect, the total throw on the thrust of only 15 cm is somewhat surprising! Similar joint orientations occur at higher levels in the cut but no additional faults were discovered anywhere else in the huge cut.

COMMON JOINTS

Common joint orientations were measured at approximately 60 stations along the Big Sandy line in the hope of finding some systems easily correlatable across the area. Joint patterns, unfortunately, varied widely from station to station or among groups of stations. A traditional approach of contrasting the several classes of fracture orientations between two nearby quadrangle areas is illustrated in Figure 8. Correlations of maxima on the Figure might be done better by mystics than by statistics!!

The difficulties in common joint correlations involve abrupt changes across boundaries of small orientation domains. These changes may occur within single road cuts with no obvious reason for the transition. Dominant joint trends may end abruptly along strike. For example, the Pikeville "canyon" is swamped with N20E trends of common joints, coal joints, joint zones, and a lineament. On strike with this trend to the NE are a series of excellent road cut exposures devoid of significant joints of this trend.

The common joints show only one reasonable persistent orientation pattern with depth. Measurements taken at different levels of a deep cut are readily correlatable in most instances. In the very deep Pikeville "canyon" cut (Figure 7) common joint orientations persist with depth but the joints themselves become increasingly rare with depth. In the core of the mountain they essentially disappear below 100 meters depth with the exception of the open joints associated with the upward splaying tiny thrust fault. A working model for the jointing in the area might include deep-level tectonic fracturing associated with minor thrusting and/or arching. These rare early joints would propagate upward to provide access for ground water and for erosion to form lineaments. Once substantial topographic relief had been established the bulk of the common jointing took place in harmony with the early fracture orientations and with the directions in which topography permitted expansion.

The patterns of common joint orientations seem to have some logic only when plotted onto a structural contour base (Figure 9). Domains of parallel or subparallel joints can be identified with axial traces of folds. South of Pikeville (Figure 9) a joint set can be seen curving in sympathy with a NE-trending axial trace, more or less parallel to the West Virginia Appalachian trends. In the area south of Prestonsburg WNW joint sets follow the area marked by the SE termination of the West Virginia fold trends. North of Prestonburg'another domain seems localized over the irregular ends of the dying folds.

COAL JOINTS

Coal joints, with spacing of a few mm to a few cm, are like the common joints in having local orientation domains which make some sense only when related to structural contour patterns (Figure 10). Coal joints show a tendency to be part of local orthogonal sets with one partner much smoother, more planar and better developed than the other, Where cleat is present the joints are more widely spaced than the cleat, are mostly parallel to it and only rarely transect it. The coal joints are typically restricted to the coal horizons and extend into the adjacent rocks only when those rocks are fine-grained siltstones of apparent mechanical similarity. In many outcrops one set of coal joints is parallel to one dominant set of common joints but in many other outcrops, the coal joints have orientations different from those of the common joints suggesting the systems differ in time and stress orientation, similar to those of Nickelsen's (1967) study of jointing in Pennsylvania coal fields.

The domain boundaries of coal jointing may be strikingly abrupt. In some coal horizons a complete change in joint orientation can occur within 30 meters with nothing evident to mark the change other than the emergence of new maxima on the equal area net as one measures and plots along the outcrop.

JOINT ZONES

Joint zones, headings, or "hill seams" in local terminology are abundant in most outcrops. Locally, they are well-known as mining hazards isolating blocks susceptible to roof falls and as highway engineering hazards permitting large sections of road cuts to collapse as huge rockfalls. These near-vertical zones of intense joint development are typically a few cm to to two meters in width with 20 cm about average. Within the zone the joints have spacings of a few cm. Typical spacings of the zones are 5 to 30 meters with vertical extents of 10-20 meters. They have a tendency to parallel common joints in an outcrop, and to have a master set somewhat parallel to the strike of the dominant topographic contour lines of a valley. Locally a suborthogonal partner may be developed. Dominant orientations change rapidly from outcrop to outcrop and the zones themselves may curve in strike.

The characteristics of these zones as' exposed in the exceptionally deep Pikeville "canyon" excavation give important clues to their origin (Figure 7). The zones are strata-bound, typically between coal horizons. They terminate abruptly against the coals but may reappear below the coal with no sign of coal disruption, Within a given mechanical horizon, typically 20-40 m thick, the zones maintain a relatively constant spacing in the mountain interior but become much more closely spaced laterally as hillside outcrop is approached, Closer to the outcrop the zones may begin to deviate from the vertical and become more parallel to the slope, The number and degree of development of the zones increases upward toward the surface with local splaying of some zones, widening of them and increasing amounts of iron staining. Downward the zones become less evident to disappear at a depth of about 100 meters below the crest of the hill or at shallower depths on either side of the crest (Figure 7).

The evidence points strongly to a non-tectonic, geomorphic origin of these features with gravity as the dominant force driving very minor lateral spreading. The process makes use of this strength anisotropy of the coal horizons, existing joint systems in the rocks and directions of easiest expansion in the local gross topography. The zones might appear as local photo-linears on aerial photographs or show local development along regional trends but it is unlikely that they are the fundamental cause of the lineaments of 10s of km length visible on satellite or radar imagery.

This geomorphic model for the origin of the joint zones provides some basis for predicting their degree of development and orientation and, as such, may be of some use in mining and highway engineering in the area.

FAULTS

The general absence of even small-scale faulting in the Big Sandy road cuts is remarkable for a region immediately in front of the Pine Mountain Thrust. In the course of five weeks' field work examining minute fracture surfaces in almost perfect exposures only four faults were observed. Two of these are illustrated in Figures 3 and 7. The other two were slickensided joint surfaces resulting from a cm or two of motion. The largest fault zone (Figure 3) has only two meters of thrust shortening and this is normal to the typical trends one might expect in the region. It is also located in a lineament of similar trend marking the approximate SW limit of the West Virginia fold trends. The other fault (Figure 7) has a thrust displacement of about 1 m along a SE dipping plane striking N20E. This trend is more in accord with the West Virginia fold strike than the Pine Mtn. motions.

The structural contours of the area (Figure 2) have irregular axial traces but their strikes in this area are also more closely parallel to the West Virginia axes than the Pine Mtn. ones. Even in the zone immediately in front of Pine Mtn. (Figure 2) the fold axial strikes meet Pine Mtn. at an appreciable angle. The evidence is suggestive that the Pine Mtn. thrusting stresses did not have an appreciable effect on this part of the Big Sandy gas field much beyond the limit of the presently mapped frontal thrusts. Instead, the minor subtle structures of the area seem much more closely related to the West Virginia fold system which seems to fade into this area and merge with transverse structures, possibly associated with the Kentucky River fault system. Nowhere were faults recognized In these outcrops as closely related to the Kentucky River system, possibly because the observations ended before reaching the Irvine-Paint Creek fault zone.

CONCLUSIONS

(1) Large scale topographic lineament systems in the Big Sandy region have some of their dominant directions at N60-70E, N20-25E, N35W, N-S, N80W. Most of these lineaments are of greater area1 extent than this portion of the Appalachians, One set may be related to the Pine Mtn. trends but in general the lineament systems might better be considered the effects of regional or subcontinental sized stress trajectories superimposed indiscriminately across the local structure of the area.

(2) Possible correlations with ground observations were found for only two of the lineaments despite excellent exposure. Both were very small-scale thrust faults with associated buckling and jointing.

(3) Vertical joint zones *or* "hill seams" are probably not the fundamental cause of the lineaments. In the deep Pikeville "canyon" cuts, the zones appear to be a geomorphic phenomenon produced by gravitative spreading of strata-bound mechanical units separated by coal seams in the uppermost 100 meters of the topography. This origin model may have some value in the prediction of these mining and highway hazards.

(4) Common joint orientation patterns persist vertically at given locations but may change abruptly in a lateral direction. Domains of parallel joint set orientations are related at least in part *to* the directions of a number of subtle fold hinges evident in structural contour maps of the area. These domains seem much more closely akin to the trends of the West Virginia fold belt and its southwestward termination than to the Pine Mtn. thrust trends.

(5) Coal joints may mimic locally the orientations of common joints but they constitute a different fracture domain which is probably of different age(s), Like the common joints, the domains of coal jointing seem linked to some of the subtle fold patterns evident in structural contour maps.

(6) In the Pikeville-Prestonsburg traverse of the Big Sandy gas field, there is surprisingly little evidence of the Pine Mtn. thrusting stresses either in density of faulted surfaces, orientation of the few faults present, strike of axial trends on structural contour maps, or of the pattern of joint domains. The dominant brittle structural features seem more closely related to the West Virginia fold trends fading southwestward into irregular transverse structural contour domains.

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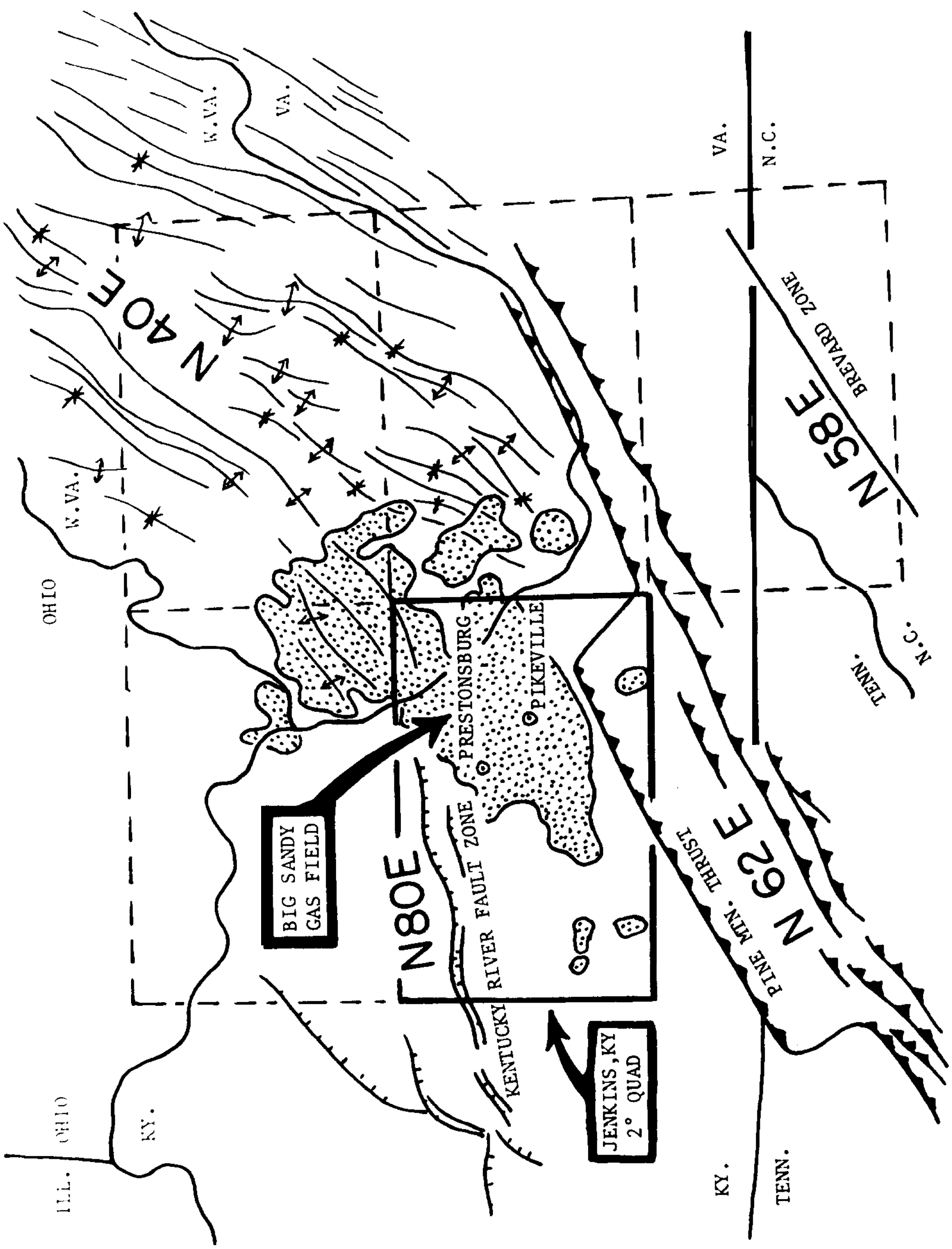
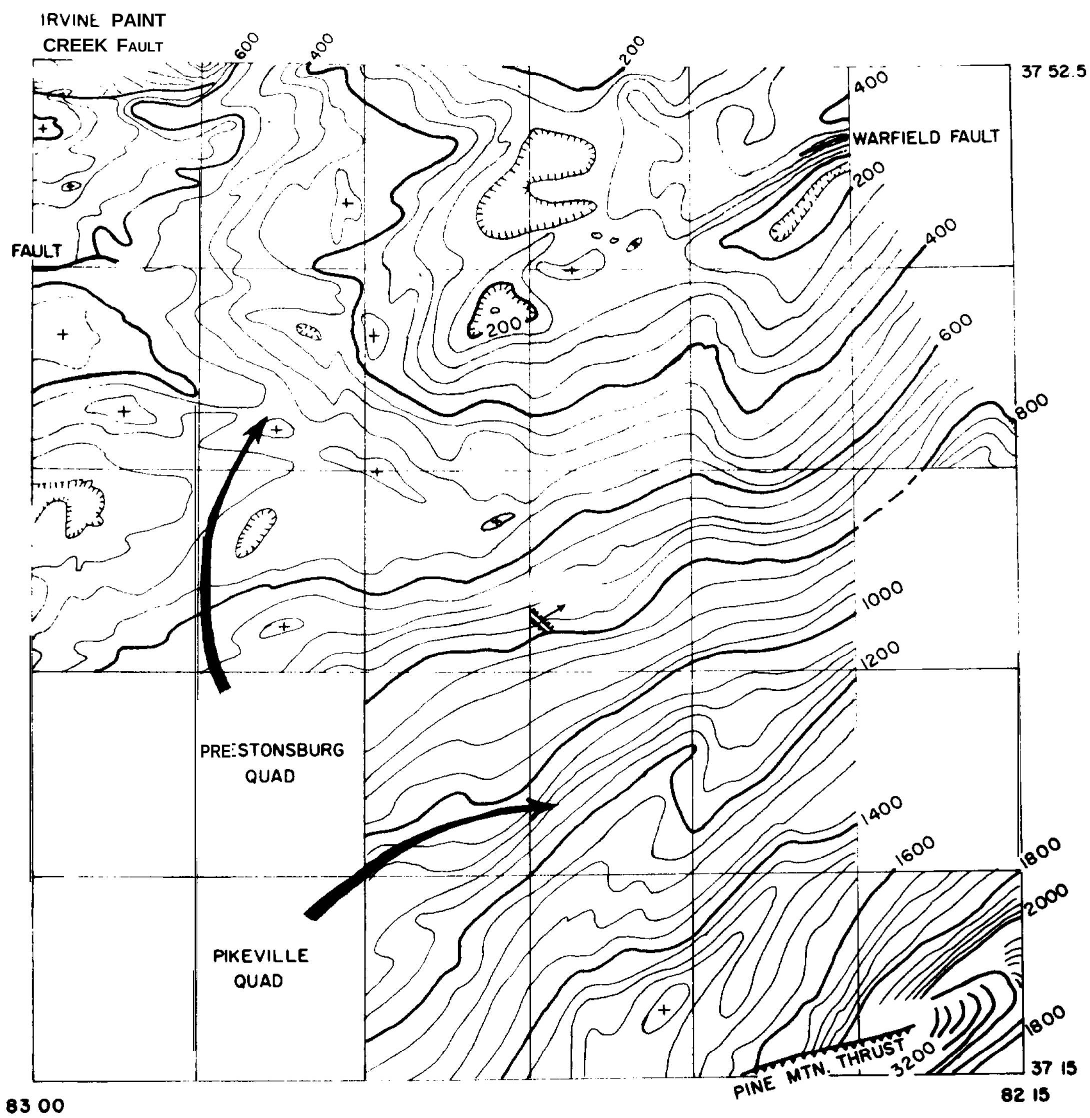


INDEX MAP TO PIKEVILLE - BIG SANDY GAS FIELD FRACTURE STUDY AREA, KENTUCKY

FIGURE 1

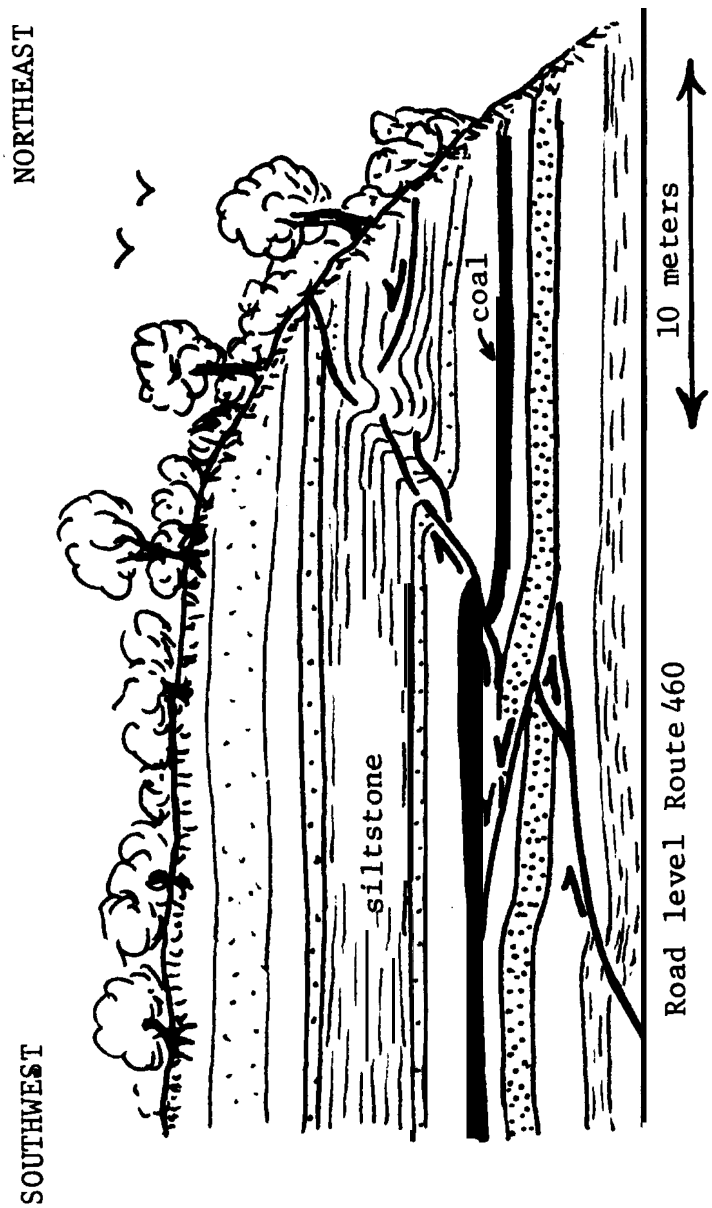


STRUCTURAL MAP OF PART OF S. EAST KENTUCKY

DRAWN ON LEVEL OF UPPER ELKHORN 2 COAL UNIT OR EQUIVALENT,

AFTER USGS QUADRANGLE MAPS

FIGURE 2



THRUSTING WITH STRIKE NORMAL TO APPALACHIAN TRENDS

Vertical beds in kink folds strike N50W. Slickensides strike N60E. Total shortening in Northeast-Southwest direction is two meters. Location: road cut on Kentucky Route 460 at the Pike County-Floyd County line.

FIGURE 3

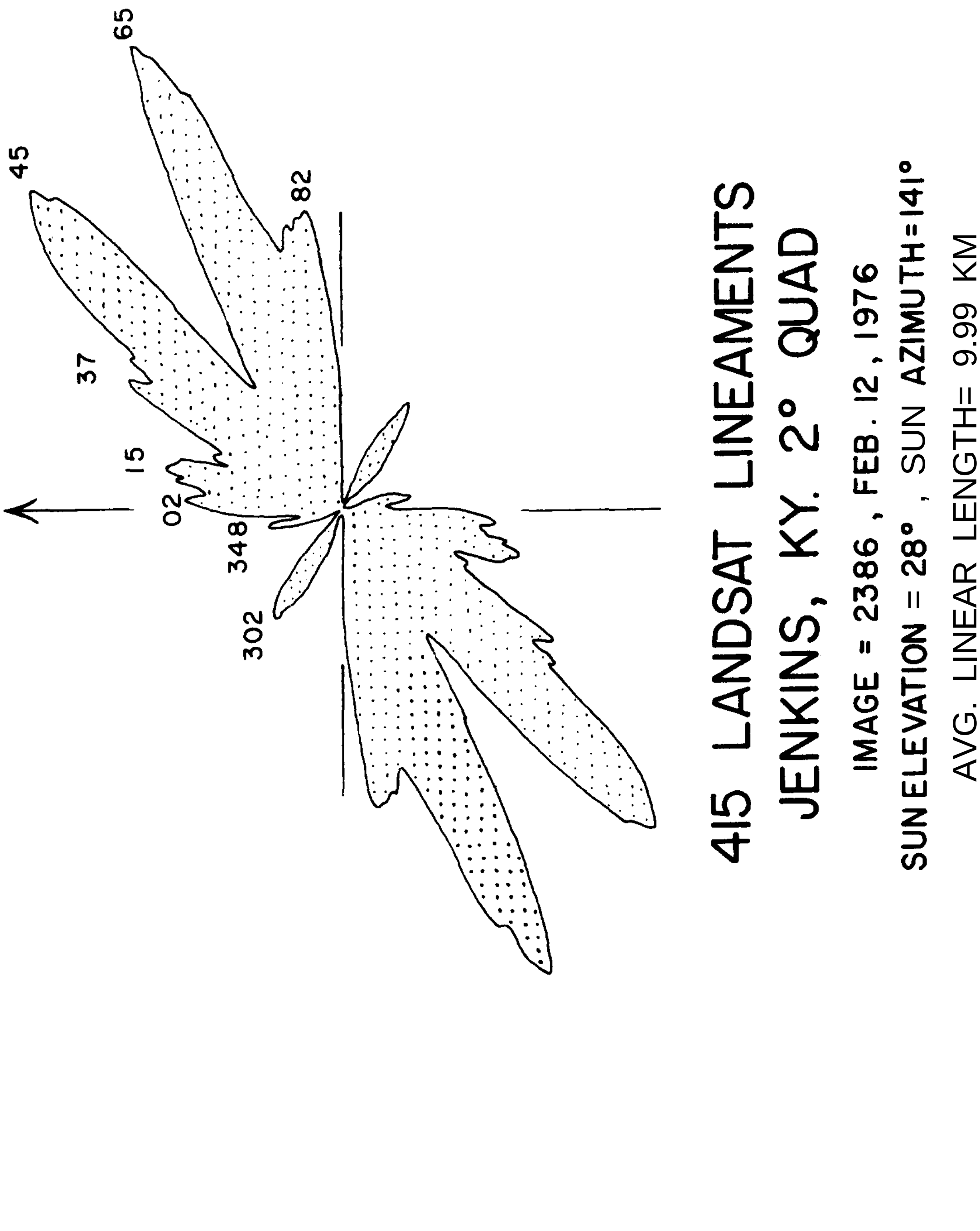

65
45
82
37
15
02
348
302
415 LANDSAT LINEAMENTS
JENKINS, KY. 2° QUAD
IMAGE = 2386 , FEB. 12 , 1976
SUN ELEVATION = 28° , SUN AZIMUTH = 141°
AVG. LINEAR LENGTH= 9.99 KM

FIGURE 4

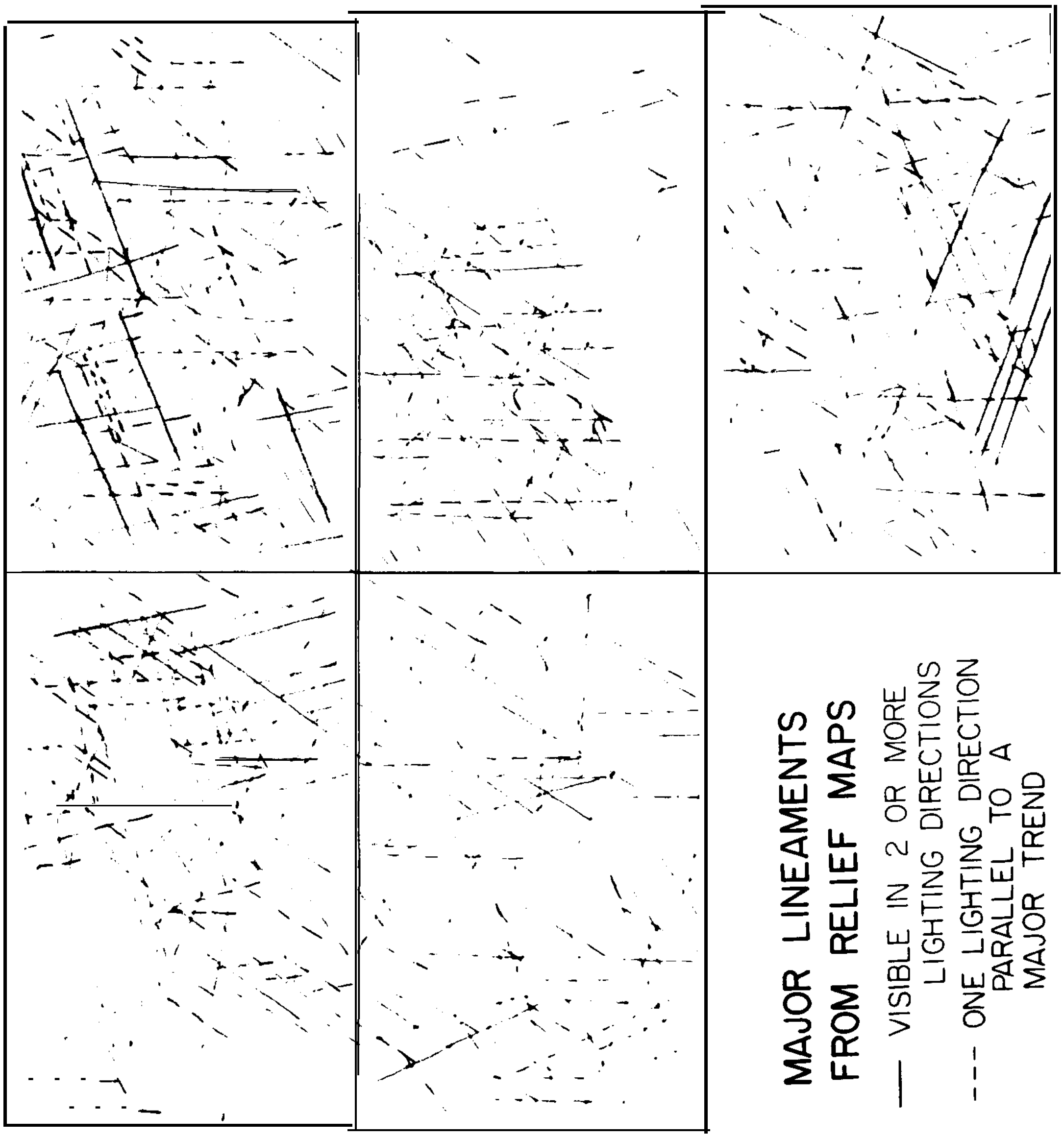

MAJOR LINEAMENTS
FROM RELIEF MAPS
— VISIBLE IN 2 OR MORE
LIGHTING DIRECTIONS
--- ONE LIGHTING DIRECTION
PARALLEL TO A
MAJOR TREND

FIGURE 5A

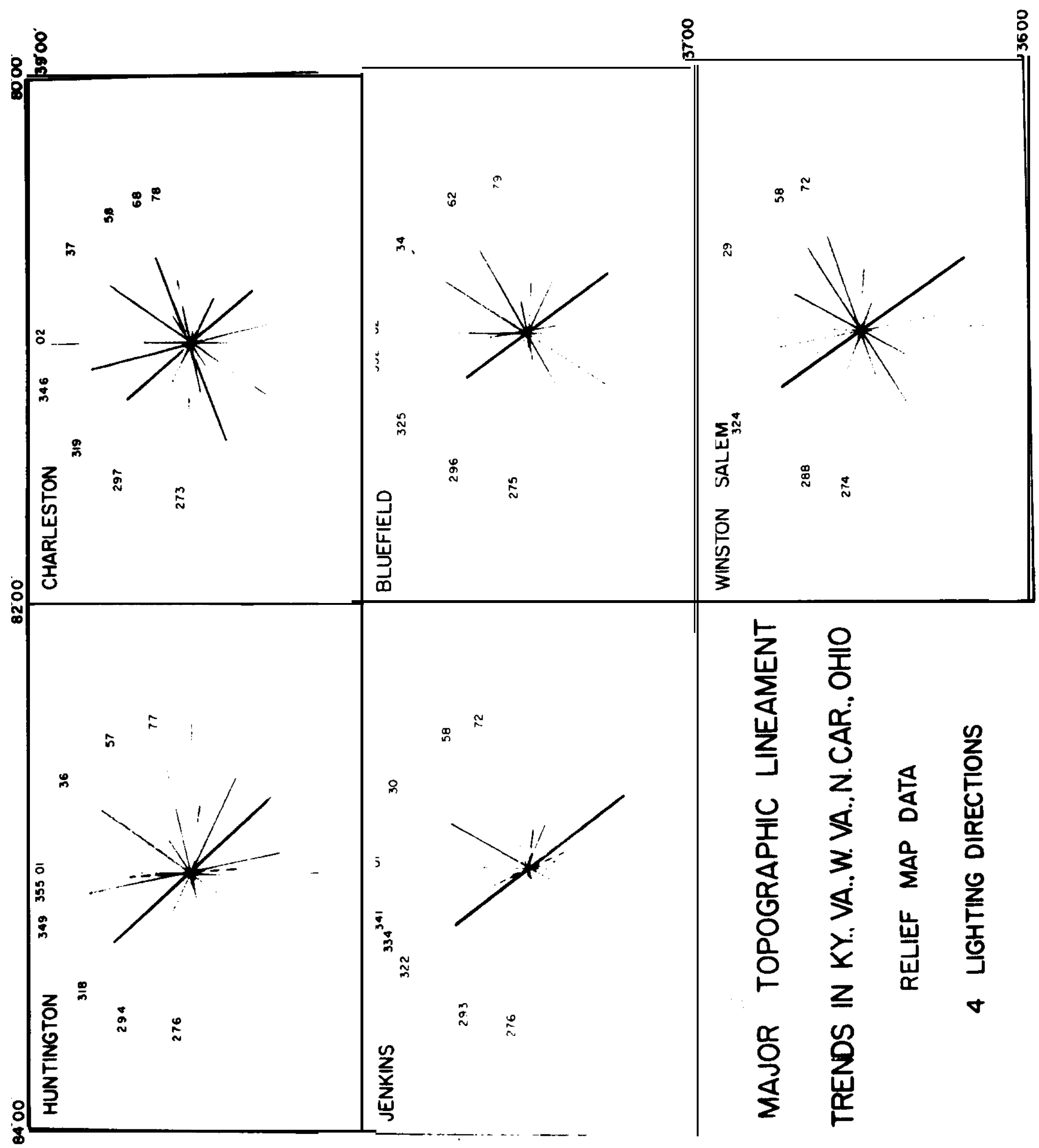

80°00'
39°00'
37°00'
36°00'
82°00'
84°00'
CHARLESTON
HUNTINGTON
BLUEFIELD
JENKINS
WINSTON SALEM
MAJOR TOPOGRAPHIC LINEAMENT
TRENDS IN KY., VA., W. VA., N. CAR., OHIO
RELIEF MAP DATA
4 LIGHTING DIRECTIONS

FIGURE 5B

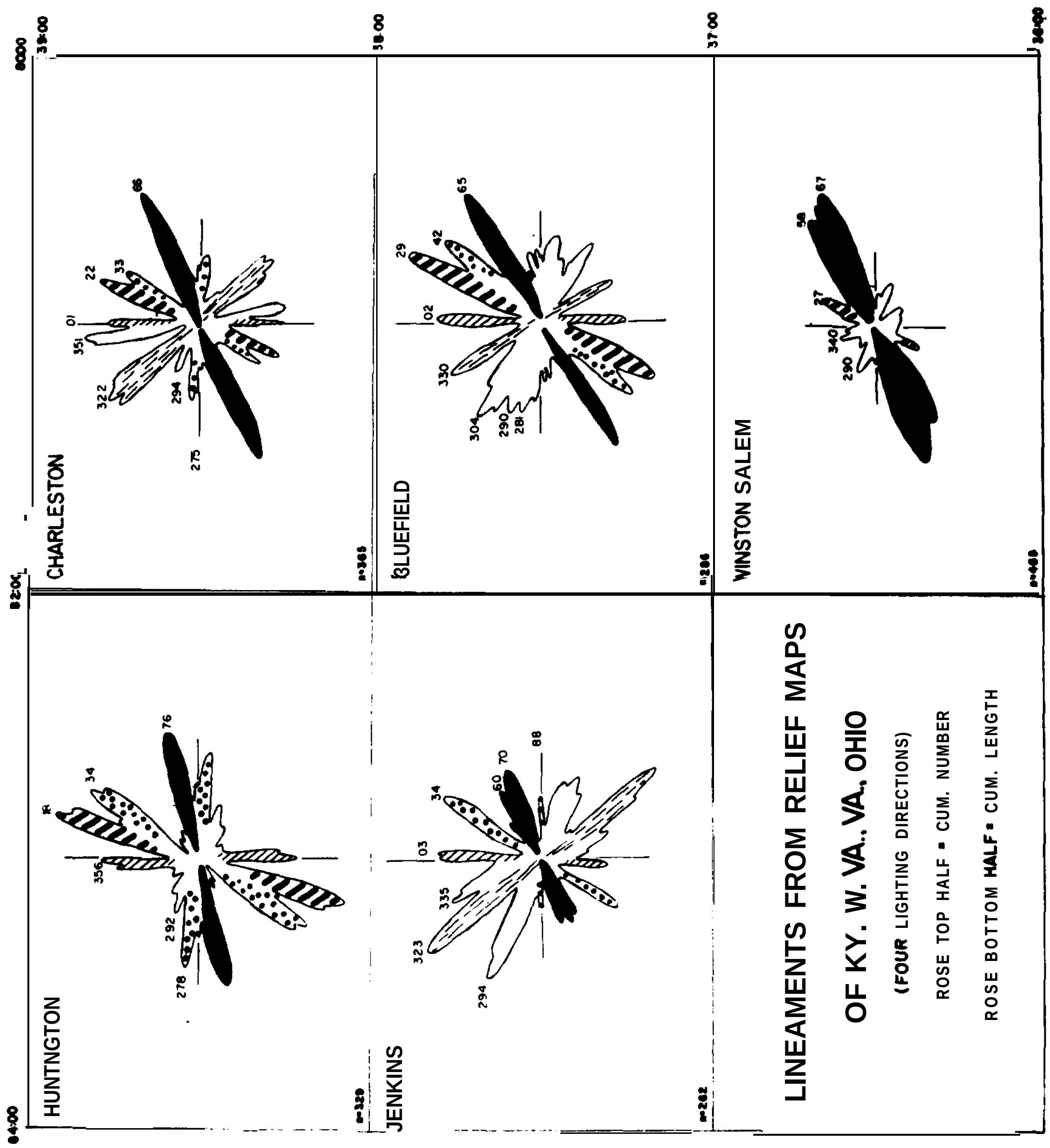
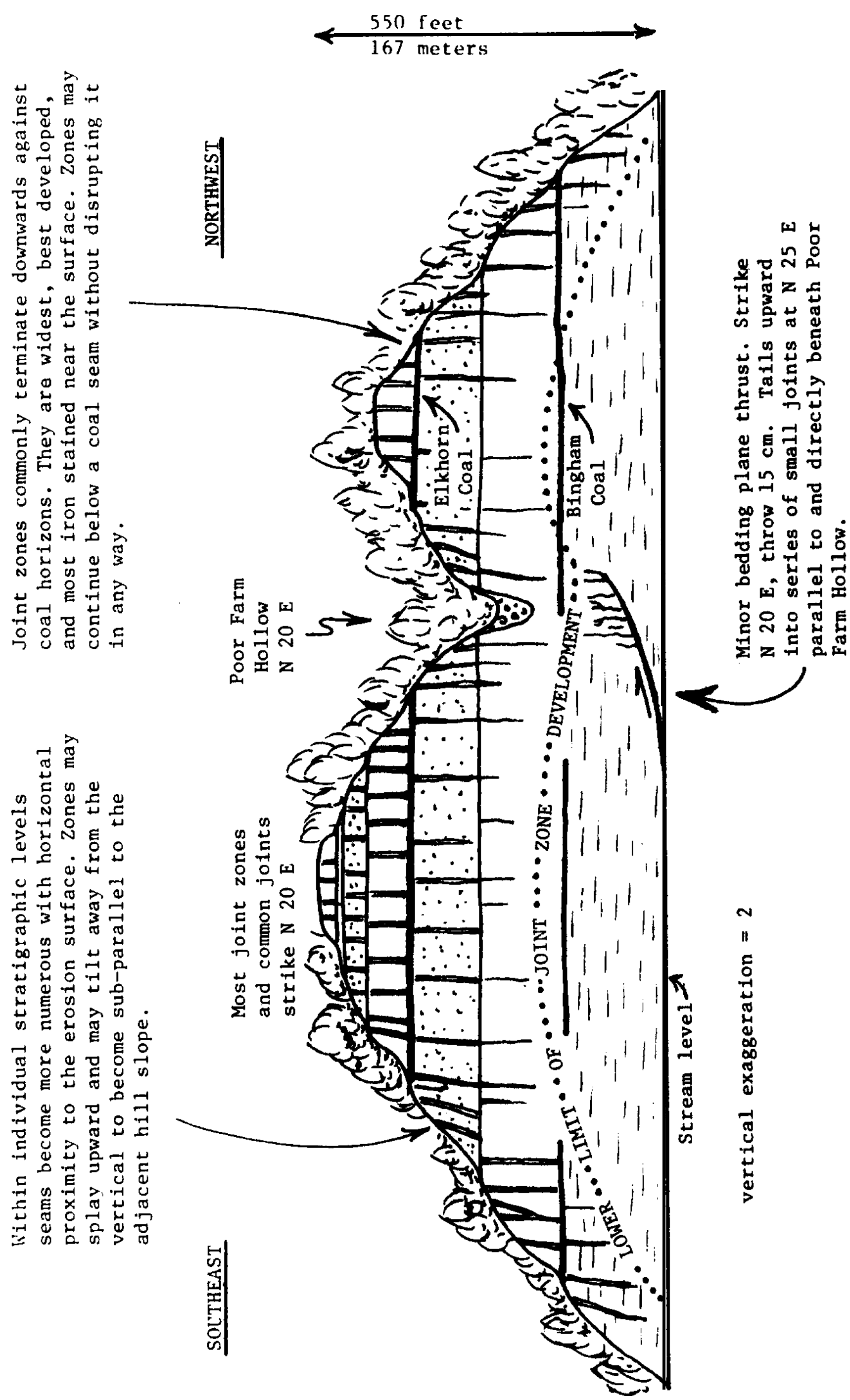

HUNTINGTON
CHARLESTON
JENKINS
BLUEFIELD
WINSTON SALEM
LINEAMENTS FROM RELIEF MAPS
OF KY. W. VA., VA., OHIO
(FOUR LIGHTING DIRECTIONS)
ROSE TOP HALF = CUM. NUMBER
ROSE BOTTOM HALF = CUM. LENGTH

FIGURE 6



JOINT ZONES OR "HILL SEAMS" AND OTHER FRACTURE CHARACTERISTICS IN THE PIKEVILLE "CANYON" EXCAVATION

FIGURE 7

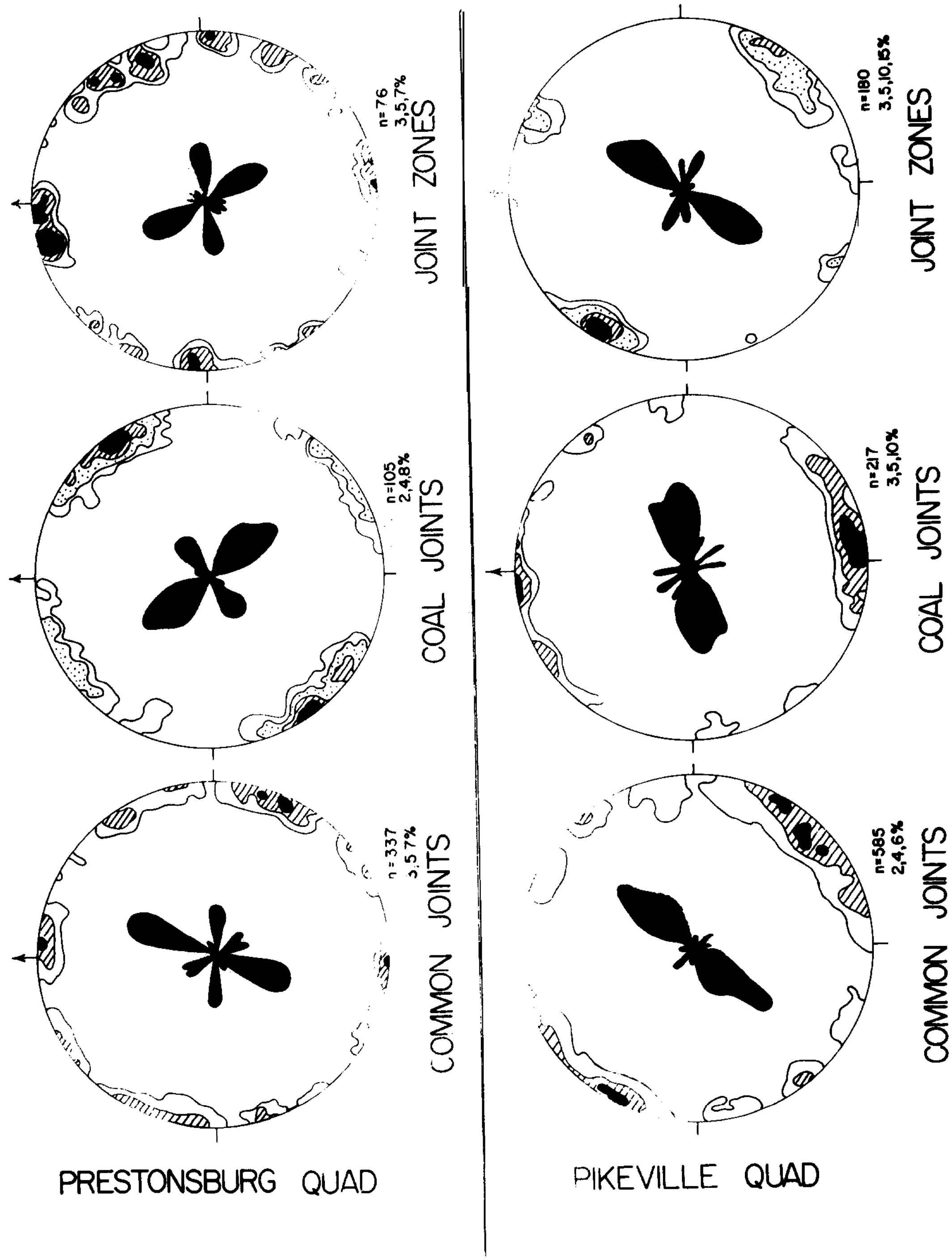

JOINT ZONES
n=76
3,5,7%
COAL JOINTS
n=105
2,4,8%
COMMON JOINTS
n=337
3,5,7%
PRESTONSBURG QUAD
JOINT ZONES
n=180
3,5,10,15%
COAL JOINTS
n=217
3,5,10%
COMMON JOINTS
n=585
2,4,6%
PIKEVILLE QUAD

FIGURE 8

COMMON JOMS OF PART OF S. EAST KENTUCKY

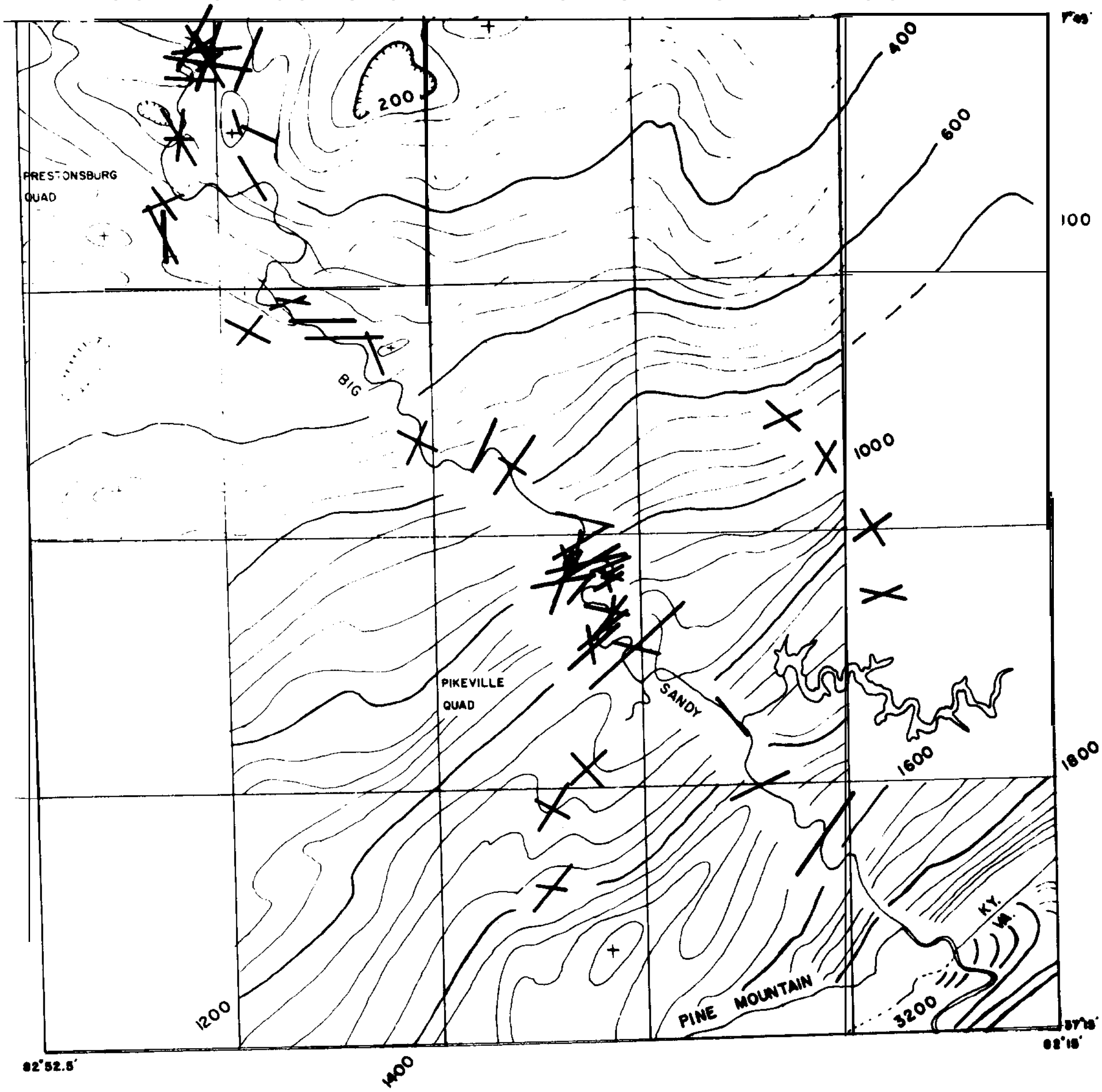


FIGURE 9

COAL JOINTS OF PART OF S. EAST KENTUCKY

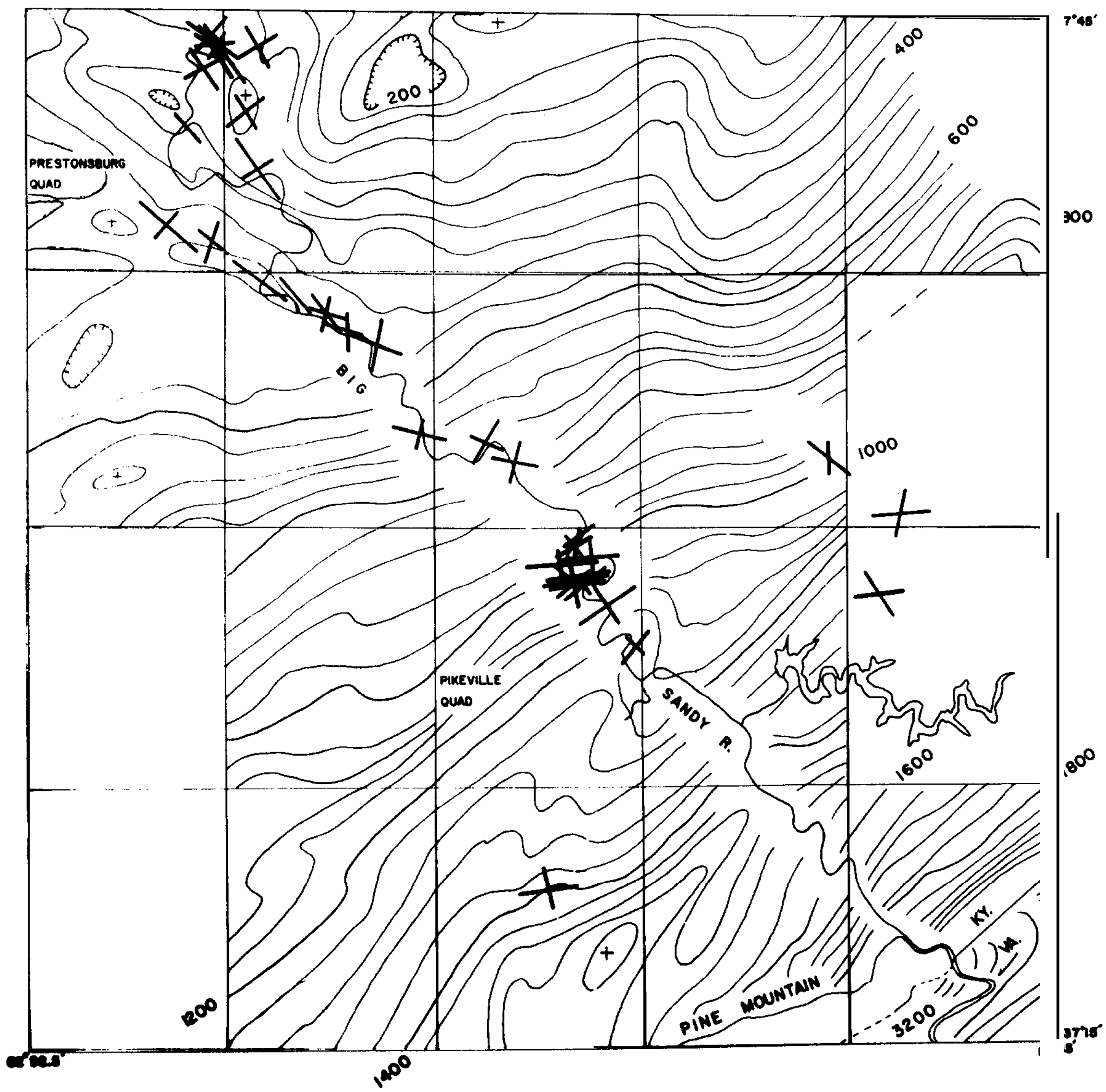


FIGURE 10

A MODEL FOR THE ORIGIN OF MINOR FAULTS BENEATH THRUST SHEETS

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ABSTRACT

The economically important Bales slice, a lower-plate thrust slice of the Pine Mountain block, Kentucky, Virginia and Tennessee, was produced by formation of a second ramping thrust fault joining the uppermost fault zone in Devonian Chattanooga Shale and an inferred lower fault zone in the Cambrian Rome Formation to the exterior of the major southeasternmost ramp of the Pine Mountain block (Miller, 1973). A force-balance mechanical model for the origin of this and other lower-plate faults shows that the net force imposed on the lower plate depends upon three quantities: 1) Δq_1, the difference in shear strength between the upper and lower fault-zone materials to the exterior of the ramp; 2) Δq_2, difference in shear strength between the ramp's fault zone and the lower fault zone; 3) bending resistance of the upper plate at the ramp. Δq_1 contributes to the state of stress in the lower plate if the upper fault zone is significantly stronger than the lower fault zone. As lower-fault-zone material is transported up the ramp, Δq_1 decreases since the same material will comprise upper and lower fault zones. Early in the development of the ramp, Δq_2 is relatively large since strong rocks are still juxtaposed on the ramp and therefore the difference in shear strength between the ramp's competent fault zone and the incompetent lower fault zone will be large. A plane strain linearly viscous model for upper-plate deformation shows that bending resistance imposes a normal stress on the ramp which is greatest at the middleoftheramp. Since bending resistance depends upon the cube of thicknesgasthe increasingly thick wedge-shaped upper plate moves into the ramp, the bending resistance and imposed normal stress will increase.

INTRODUCTION

Many major thrust faults are not simple fractures but a system of faults that have had different histories on different branches. One of the best known examples is the Pine Mountain and subsidiary faults of parts of Kentucky, Virginia and Tennessee of the southern Appalachians (Fig. 1). In the central part of the block, the fault which eventually outcrops as one fault is actually two faults beneath Powell Valley anticline; the uppermost one is called the Pine Mountain fault and the lower the Bales fault (Fig. 2, section bb'; compare Harris, 1970 and Miller, 1973, Fig. 1). The slab of lower-plate rock between the two faults, the Bales Slice, has moved about 4.8 km (Harris, 1970, p. 163), or perhaps less, in concert with the upper Pine Mountain block proper.

Bow did the Bales Slice and other lower-plate slices like it come about? Why did the Bales fault ramp about 4 km beyond the ramp of the Pine Mountain Fault? There are clearly two parts to its evolution, though not necessarily separated in time. First, for the Bales Slice to move, the lower Bales fault had to form. (Miller [1973, Fig. 1] has the sole fault from which the Bales fault branched actually extend beneath the entire Pine Mountain block.) The Bales fault consists of a bedding Plane portion as well as a ramp along which the Bales fault climbs from the Cambrian Rome Formation to the Devonian Chattanooga Shale. Second, movement occurred on the Bales fault, moving the slice up its ramp, and as a result warping the Pine Mountain fault above the slice.

Both the formation of the Bales fault and its movement were controlled by the stresses imposed on the lower plate by those rocks above. In order to focus on how these forces arose, a simple force-balance model is developed below. A complete mechanical model was not attempted. Instead, this model is developed to elucidate the role of the various boundary stresses through time in raising the state of stress in the lower plate. Therefore, the lower plate is considered to be a rigid block bounded above and below by fault zones as well as to the southeast by the ramp of the Pine Mountain fault and to the northwest by a potential area of failure a distance L from the top of the ramp of the Pine Mountain fault. The specific questions to be addressed are, 1) what was the role of the relative strengths of the fault zones, the related problem of, 2) how does this affect the distance from the top of the ramp at which the Bales fault ramps and finally, 3) what was the role of the bending stresses imposed by the upper plate on the lower plate in raising the level of stress in the lower plate?

THE MODEL

Faults in the lower plate in the vicinity of major ramps could have formed at the same time as ramps as a result of the same processes; stress transmitted from the interior of the mountain belt caused failure in those rocks above the sole fault. However, the possibility also exists that lower-plate thrusts result from subsequent movement of the upper plate where large stresses are applied to the lower plate through the upper plate's fault zone. Both possibilities will be explored here.

The model is shown in Fig. 3. The heavy upper black line is the fault zone between the upper and lower plates whereas the lower lighter line is the horizon in which the sole thrust is developed. For the Pine Mountain block, the upper fault is the Pine Mountain fault, ramping from the Cambrian Rome to Devonian Chattanooga Shale, whereas the lower fault is the one inferred by Miller (1973) to lie in the Cambrian Rome Formation beneath the entire Pine Mountain block. The lateral length of the Ramp is ℓ whereas the distance from the top of the ramp to the potential area where the ramp of the lower fault may begin is given by L.

Summing forces in the horizontal (actually regional dip) direction yields

$$F = F_{qr} + F_{qa} + F_{P} - F_{qb} \tag{1}$$

where F is the force with which the lower-plate rocks some distance L beyond the top of the ramp must exert to maintain equilibrium, F_{qr} is the force due to the shear stress on the ramp caused by movement of the upper plate, F_{qa} is the force due to shear stress on the upper fault zone beyond the ramp, F_{qb} is the force due to shear resistance to movement of the lower fault zone, F_p is the force imposed on the lower plate by the bending resistance of the upper plate at the ramp.

The expressions F_{qa} and F_{qb} are straightforward. They are:

$$F_{qa} = q_a L \tag{2}$$

$$F_{qb} = q_b L \tag{3}$$

or just the shear strengthsof the respective fault zones (q_a and q_b) which arise due to movement times the length of the fault zone.

F_{qr} and F_p, however must be found by summing the resolved shear and normal bending stress vectors (tractions) on the ramp in the horizontal direction (Fig. 4):

$$F_{qr} = \int_{o}^{2S_o} q_r \cos(\theta)\, ds \tag{4}$$

$$F_P = \int_{o}^{2S_o} p \sin(\theta)\, ds \tag{5}$$

S_o is half the arclength of the ramp, Θ the ramp dfp and ds a small increment of arclength, Since $2S_o$ = ll, to a first approximation

$$F = \Delta q_1 L + 2\Delta q_2 S_o + \int^{2S_o} p\sin(\theta)ds \quad (6)$$

where

$$\Delta q_1 = q_a - q_b \quad (7a)$$

$$\Delta q_2 = q_r - q_b \quad (7b)$$

Therefore, the force exerted on a vertical surface joining the upper and lower fault zones a distance L from the top of the ramp (Fig. 3) is approximately equal to the product of the distance, L, and the difference in strengths of the upper and lower fault zones plus the product of the parallel to regional dip component of the force due to the difference in strength of the ramp fault zone and the lower fault zone and the length of the ramp, plus the force due to bending resistance of the upper plate on the ramp.

DISCUSSION

There are three possibilities for the role of the relative strength of the upper and lower fault zones: beyond the ramp (1st term, Eq. 6) the upper fault zone (q_a) is 1) equal in strength to, 2) weaker than, or 3) stronger than the lower fault zone (q_b). For case 1), the first term of (6) plays no role in determining the net traction F/d at any distance L from the *top* of the ramp (Fig. 3); this simple model would predict in this case no reason for a ramp in the lower fault zone to take place at any distance from the major ramp. If the upper fault zone were weaker, however, the first term in (6) is negative and an additional force must be added at the ramp (second or third term of Eq. 6) to overcome the resistance of the lower fault. Under the last case, movement of the upper plate imposes a force that exceeds the resistance of the lower fault zone and therefore Aq, is greater than zero so that the traction F/d is larger further away from the top of the ramp. In this case, Aq, has some control on the distance from the top of the ramp at which the net traction is high enough to favor failure.

Each of these cases finds some support from either experimental data or geological argument. The experiments of Handin and Hager (1957) show that those shales they tested have similar shear strengths (one half the ultimate strength) both normal and parallel to bedding, implying Aq, would be small. Wilson (1970) however, concludes that shales with a higher degree of compaction are stronger. Therefore, the lower fault zone of Fig. 3 would be stronger and Aq, would be negative. Finally, the possibility exists that shale becomes stronger with large amounts of shear as a result of processes such as loss of fluid pressure or strain hardening. In this case Aq, would be positive and movement of the upper plate would contribute to the state of stress in the lower plate. The fact that the Bales fault ramped a distance away from the top of the ramp supports this last case.

Δq_2 must be quite large during the initial stage of ramping of the upper plate when the competent upper plate rocks are still in contact with lower plate rocks across much of the ramp. It is most likely at this stage, then, that the second term in (6) plays the largest role in increasing F/d. As a numerical example, if the stratigraphic distance between the upper and lower fault zones is 2000 m (Rome - Chattanooga distance) and if the arclength ($2S_o$) of the upper plate ramp is about 11.3 km, a compressive traction of 4 kb (0.4 GPa) will be imposed on the 2000 m section if $q_r - q_b = F/2S_o$ = 4 kb * 2 km/11.3 km = 708 b (70.8 MPa), or if the value of Δq_2 is 1/6 that of the block. Thus, in the early stages of movement, when q_r is significantly larger than q_b, stresses are higher than perhaps at any other time; Aq, decreases with motion of the upper plate because weak fault-zone shale is brought up by the upper fault and separates the upper and lower plates (Fig. 2). Δq_2 plays an early role in the formation of lower-plate faults.

The effect of bending resistance, the last term in (6) increases with ramp dip, upper-plate thickness and fault-zone strength. The normal stress p is given by (Wiltschko, 1978):

$$p = \theta_o \beta[\frac{P}{2} \sin(\beta s) - \eta_t h(1-A)V\beta\cos(\beta s)] + \frac{\Omega\xi}{\delta s} \quad (8)$$

where

$$\beta = \frac{\Pi}{S_o} \tag{9a}$$

$$A = \frac{1}{1 + \frac{h^2\beta^2}{6}\ \frac{\eta_n}{\eta_t}} \tag{9b}$$

V is the velocity of movement of the upper plate, θ_o the maximum dip of the ramp, P the force driving the thrust sheet forward (in the s-direction); η_n and η_t are the compression and shear viscosities (Biot, 1965), h the thickness of the upper plate and $\Omega\xi/\delta s$ the gravitational term. P is in turn given by

$$P = 2S_o q_r + JS_o + \int_o^{2S_o} \frac{\partial\Omega s}{\partial s}\, ds + P_a \tag{10}$$

$$J = \frac{1}{12}\, \eta_n h^3 V\theta_o^2 \beta^4 A S_o \tag{11}$$

The third term on the right in (10) is the gravitational term and P_a is the force that must be exerted at the top of the ramp to move all those rocks in front of the ramp. For typical Pine Mountain block parameters (Wiltschko, 1978) F_p, combining (8), (9), (10) and (11) in (5), can be of the same order of magnitude as F_{qr}. See Fig. 5.

The effect of the ramp dip, θ_o may be seen in(ll). A larger ramp dip for the same ramp length results in a more abrupt ramp and thus a larger normal stress on the lower plate. The thickness of the upper plate appears as the cube in (11) and thus in P and p. As the increasingly thick wedge-shaped upper plate moves into the ramp, the bending resistance will increase, perhaps significantly. Therefore, p will increase with displacement of the thrust sheet since the stratigraphic wedge of which it is a part increases in thickness in the direction from which the thrust sheet moves.

The normal stress on the lower plate, p, is also increased through P by q_r and q_a because P must be increased to overcome larger fault-zone strengths at a constant displacement velocity. Therefore, if shales become stronger with deformation, not only will Aq, increase, but also p, if the same velocity of movement of the upper plate is maintained. Consequently, other than the initial stage of ramping of the major thrust fault, the other most likely time of formation of the Bales slice and other subsidiary lower plate splays is later in the major thrust sheet's development when either the upper fault zone gains strength such as by loss of fluid pressure and/or as an increasingly thick upper plate imposes an increased normal stress due to bending on the ramp.

SUMMARY

Since the reason why thrust faults ramp where they do is poorly understood, the possibility exists that the Bales fault formed early by the same processes that formed the main southwest ramp of the Pine Mountain block; formation of and movement on the Bales fault could have taken place concurrently with that of the Pine Mountain fault itself.

However, there are several potential processes which take place with movement that could also give rise to lower-plate faults:

1) High shear stresses across the major fault early in its movement history;
2) Increase in strength of the upper fault zone, whether by loss of fluid pressure or work hardening of the materials in the fault zone;
3) Bending resistance of the upper plate.

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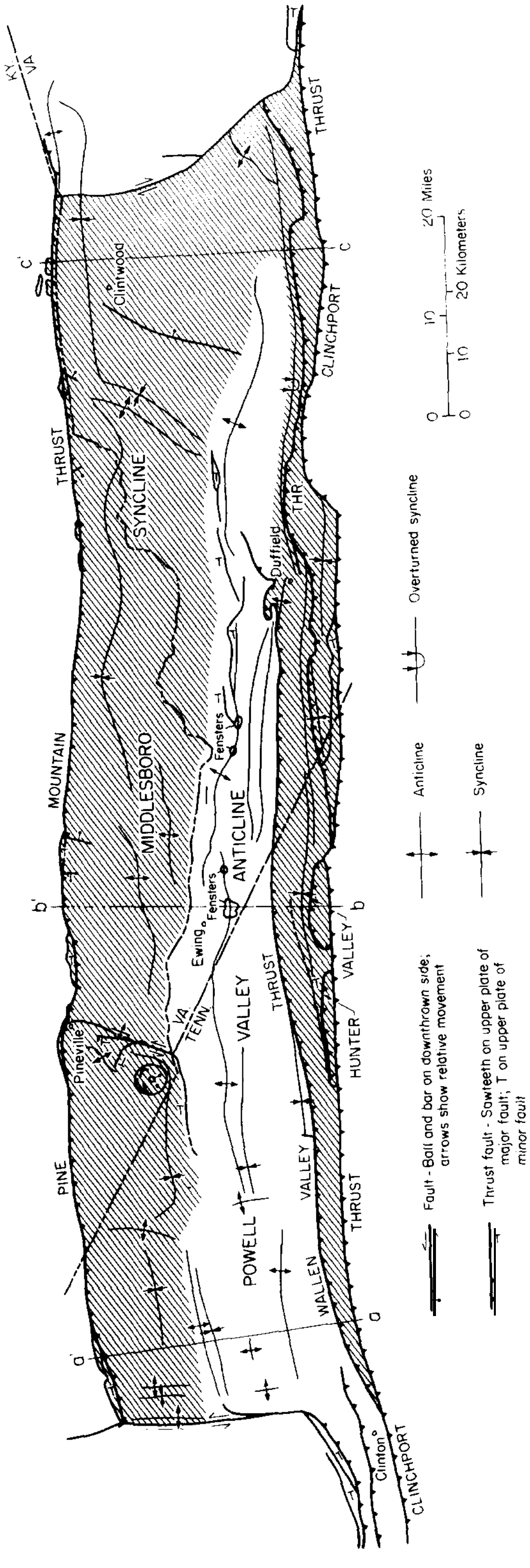

Figure 1. Pine Mountain block. The two main structural features of the Pine Mountain block are Powell Valley anticline (unhatched) and Middlesboro syncline. Sections a-a', b-b', and c-c' are shown in Figures **2a,** 2b and **2c,** respectively. The Pine Mountain block trends approximately **N60E.** From Wiltschko (1978).

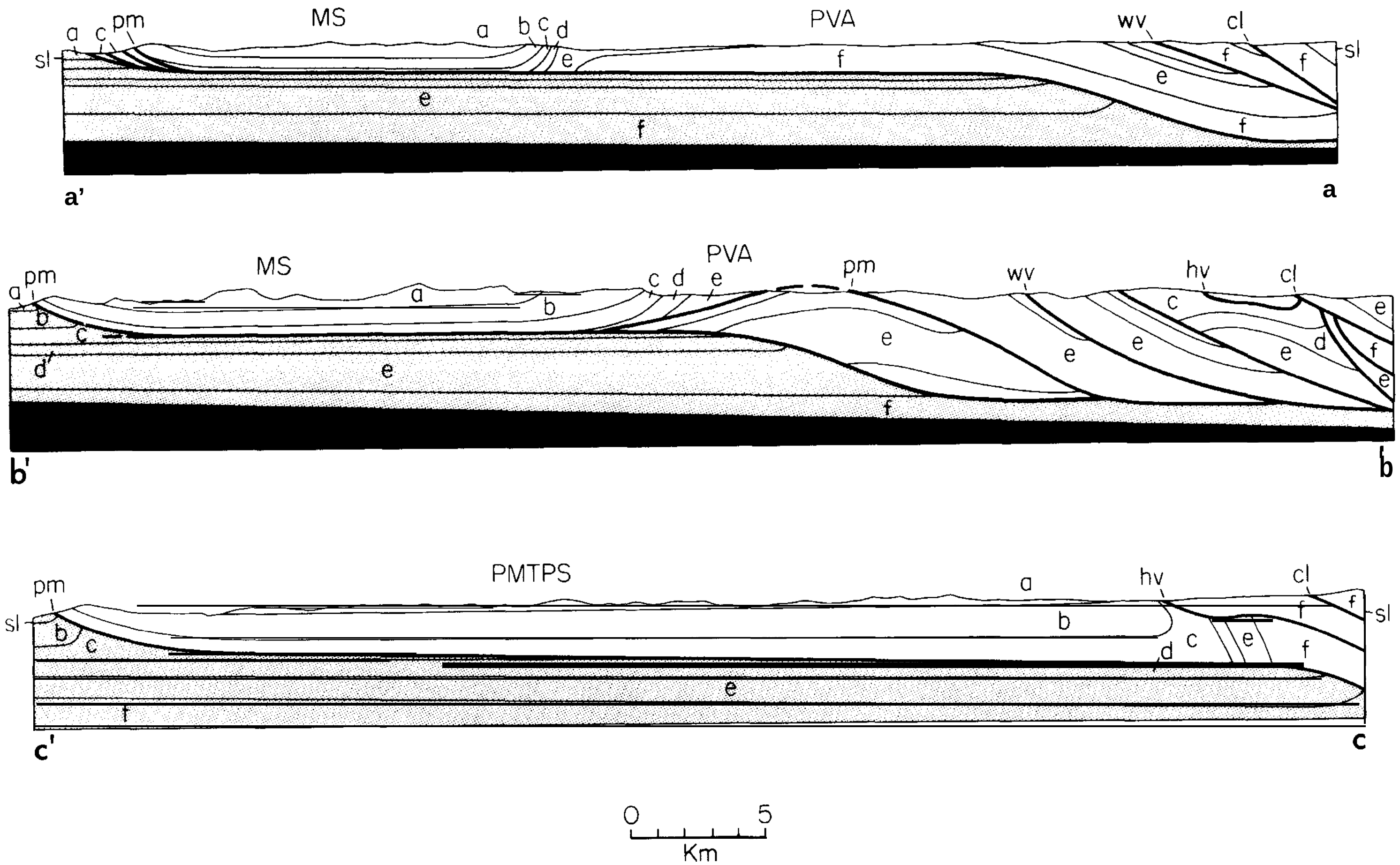

Figure 2. Cross sections through the Pine Mountain block. Each cross section (located on Figure 1 as a-a', b-b' and c-c') contains two ramps where the Pine Mountain fault climbs to a higher stratigraphic position in the direction of transport and one flat where the fault is at regional dip. Section b-b' in addition shows the location of the Bales slice caught between the warped Pine Mountain fault (pm) and lower Bales fault beneath Powell Valley anticline (PVA). Symbols: pm, Pine Mountain fault; MS, Middlesboro syncline; PVA, Powell Valley anticline; PMTPS, Pine Mountain thrust plate syncline (Harris, 1970); wv, Wallen Valley fault; hv, Hunter Valley fault; cl, Clinchport fault; a, Pennsylvanian Breathitt group and younger; b, Mississippian and Pennsylvanian coarse clastics; c, Silurian through Mississippian shales and siltstones, including the Devonian Chattanooga shale; d, primarily Trenton limestone; e, primarily Knox Group; f, Rome and Conasauga formations. The solid black areas are basement and the stippled areas are the lower plate.

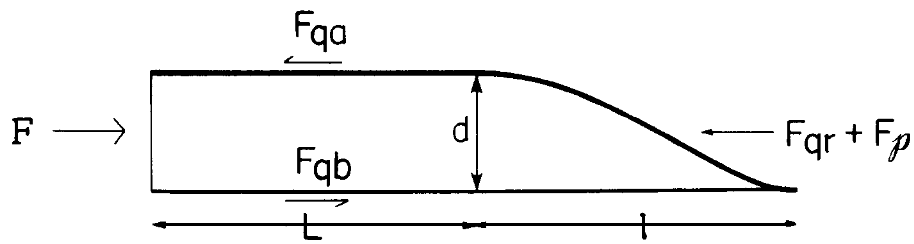

Figure 3. Force-balance model. Forces which act on upper fault zone (heavy black line) and lower fault zone (lower line) which must be resisted by strength of lower plate rocks. d is the height of the ramp, L the length of the flat, and 1 the lateral length of the ramp.

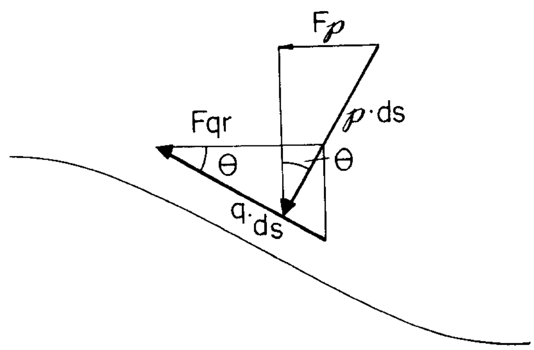

Figure 4. Forces on the ramp.

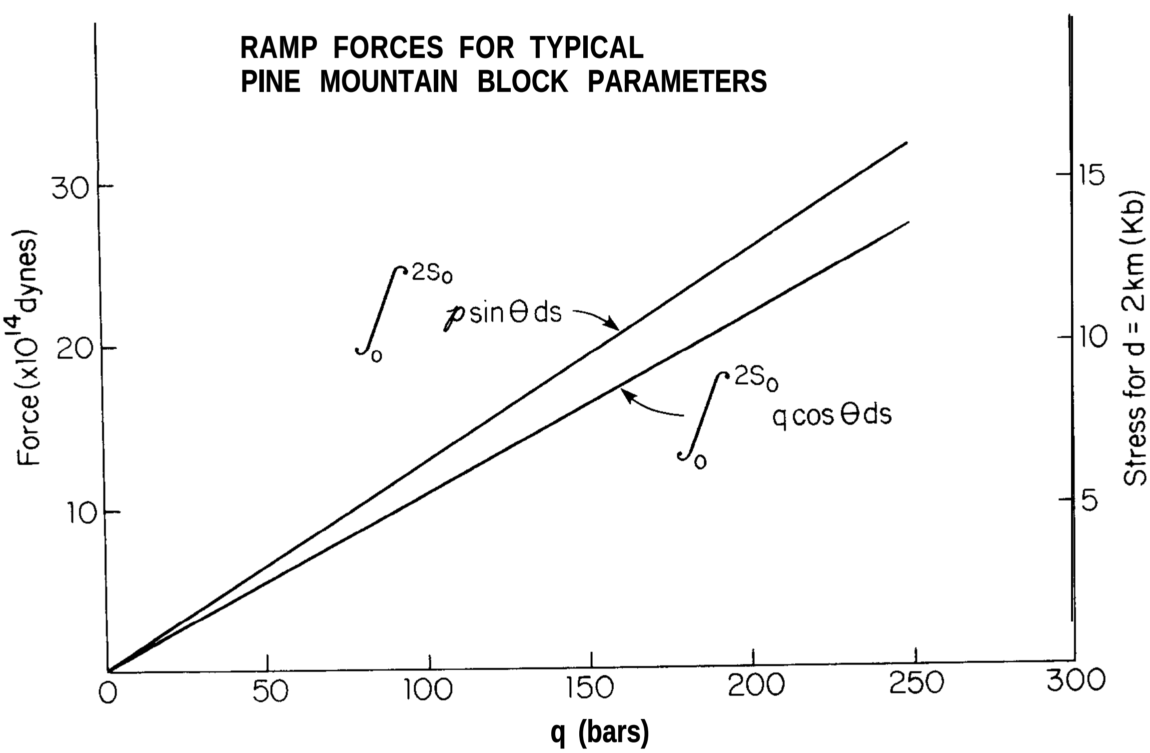

Figure 5. Forces imposed on ramp due to bending (top line) and shear (bottom).

FRACTURES RELATED TO MAJOR THRUSTS--POSSIBLE ANALOGUES TO TECTONICALLY FRACTURED CHATTANOOGA SHALE IN TENNESSEE

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ABSTRACT

In places southern Appalachian thrust blocks, both hanging walls and footwalls, are extensively fractured adjacent to the major faults. Fractures are commonly a mixture of extensional and contractional faults generated by tensional, compressional and shear stresses. Detachments are closely controlled stratigraphically, especially by coal beds in the Cumberland Plateau and by shale units in the Valley and Ridge. In the southern Cumberland Plateau and southwestern Valley and Ridge of Tennessee the Pennington-Gizzard stratigraphic interval is the preferred zone for upper level decollement. In the northern Cumberland Plateau and northwestern Valley and Ridge of Tennessee the Chattanooga Shale commonly localized upper level thrusts. Near and below the Pine Mountain block in Tennessee and Virginia, and in the footwall of the Clinchport thrust near Sneedville, Tennessee, the Chattanooga is tectonically fractured. Seismic surveys in the northwestern Tennessee Valley and Ridge suggest that these fractured zones extend beneath major thrusts, where they may be suitable traps for natural gas.

INTRODUCTION

The Tennessee Division of Geology, under contract (No. EY-76-C-05-5196) with the U.S. Department of Energy is attempting to determine if tectonically fractured Chattanooga Shale lies beneath thrusts of the southern Appalachian Valley and Ridge province (Fig. 1). The Chattanooga Shale is potentially both a source and a reservoir for natural gas.

Analyses of several samples collected by the Division of Geology in the northwestern Valley and Ridge indicate that the Chattanooga would yield between 3.0 and 4.3 gallons per ton of oil. A diamond drilling program currently in progress (April 1978) is designed to provide additional stratigraphic and chemical data for the Chattanooga Shale in eastern Tennessee.

Devonian black shales produce gas in parts of Ohio, West Virginia, eastern Kentucky and southwestern Virginia (Fig. 1). The shale yields natural gas on the Pine Mountain block in southwestern Virginia (Young, 1957). A thin zone of intensely deformed beds near the base of the Devonian shale section contains gas under high pressure. This zone, known as the blowout zone, serves as the glide plane for part of the Pine Mountain block. Similar tectonically fractured beds were tested by wells in eastern Tennessee (Fig. 1). In some of the wells gas was encountered under high pressures in deformed Chattanooga Shale. Other wells containing fractured

Chattanooga Shale were dry, and presumably any previously existing gas had escaped.

The extent of tectonically fractured Chattanooga Shale in the Tennessee Valley and Ridge is unknown. It is likely that the Chattanooga lies beneath much of the Pine Mountain block (Harris, 1970) and perhaps over a considerable area beneath thrust sheets in the northern part of the Tennessee Valley and Ridge. Tectonically fractured Chattanooga Shale exposed near Sneedville (Fig. 2) may be related either to a bedding thrust that extends in the shale beneath the younger Mississippian rocks of Newman Ridge or to footwall deformation along the Clinchport fault (Harris and Miller, 1963; Harris and Mixon, 1970).

The purpose of this paper is to summarize some of the existing data concerning fractures related to thrust faulting Tennessee, because it is anticipated that the Chattanooga Shale might contain similar structures where it is in tectonically similar situations at depth.

FRACTURES RELATED TO OVERTHRUST FAULTING

Harris and Milici (1977) developed a general model for mesoscopic deformation observed in thrust sheets in the southern Appalachians (Fig. 3). In some places they found that thrust sheets above decollement could be divided into a lower broken formation zone and an upper zone of fracture, which graded vertically into relatively undeformed strata.

In general, broken formation zones of thick tectonic breccia consist of complexly fractured and folded strata. Sandstones appear crushed and are cut by many small fractures that in places obliterate bedding. Shales are tectonically thickened and thinned by numerous fractures. Shale beds are commonly intricately folded and in places contain fracture cleavage and kink bands. Broken formation zones are exposed in a number of places along the upper level decollements of the Tennessee Cumberland Plateau. However, broken formations are exposed in only a few places on the hanging walls of thrusts in the Valley and Ridge, where the zone is presumed to have been generally abandoned at depth along the toes of tectonic ramps (Harris and Milici, 1977). A broken formation zone was described in the Blue Ridge by Milici (1978a), but there is overlain by folded rather than fractured beds.

The zone of fracture contains a mixture of extensional and contractional faults described by Harris and Milici (1977, p. 25-26) as varieties of splay thrusts and normal faults (Fig. 3). This mixture of extensional and contractional faults is widespread in the hanging walls of southern Appalachian thrust sheets. In addition to the localities described by Harris and Milici (1977) in the Cumberland Plateau and Valley and Ridge of northern Tennessee and southwestern Virginia, Milici (1978 b, c) described similar structures in the Whiteoak Mountain synclinorium and in a fault slice along the toe of the Blue Ridge.

The mixture of tensional and compressional structures is related to the development of differential stresses, and can be explained in terms of thrust sheets moving over obstacles or irregularities in the decollement zone during deformation. Computer generated cross sectional models of stress fields in thrust sheets, which show that zones of tension develop in the hanging wall near obstacles at the base of the model, may in fact be analogues to stress fields that developed in nature (Advani, GangaRoo, Chang, Dean and Overbey, 1977) (Fig. 4).

AREAS POTENTIALLY FAVORABLE TO THE DEVELOPMENT OF TECTONICALLY FRACTURED CHATTANOOGA SHALE IN EASTERN TENNESSEE

Major Appalachian thrusts rise from decollement in Cambrian shales, first cutting diagonally across overlying competent carbonate sequences and then flattening as upper level decollements in shaly formations such as the Athens or Sevier (Ordovician), Rockwood (Silurian), Chattanooga (Devonian-Mississippian), Pennington (Mississippian) or Gizzard (Pennsylvanian).

The Chattanooga serves as an effective upper decollement glide zone in Tennessee and southwestern Virginia where it is greater than 50 feet thick (Fig. 5). To the southeast, where the Chattanooga is thin, thrust faults bypass the Chattanooga on the ramps, refracting instead into the bedding of the overlying Pennington or Gizzard.

It is probable that deformation in the bedding of the Chattanooga Shale beneath the Pine Mountain block is similar to deformation observed elsewhere in surface decollements, i.e. the blowout zone described by Young (1957) is a broken formation that would be overlain by a zone of fracture (Fig. 6).

Seismic cross sections prepared by Geophysical Service Inc. were made primarily to determine the location of the Chattanooga Shale beneath Valley and Ridge thrust sheets (Tegland, in press; Milici, Harris and Statler, in preparation). Two lines, designated K-l (South)-TC-1 and TC-2 extend en echelon from the Pine Mountain block to the Blue Ridge (Fig. 1). Interpretations of segments of these lines which cross the Saltville fault suggest that there is a sizeable area where tectonically fractured Chattanooga Shale may exist at shallow depths beneath the thrust sheet (Fig. 7).

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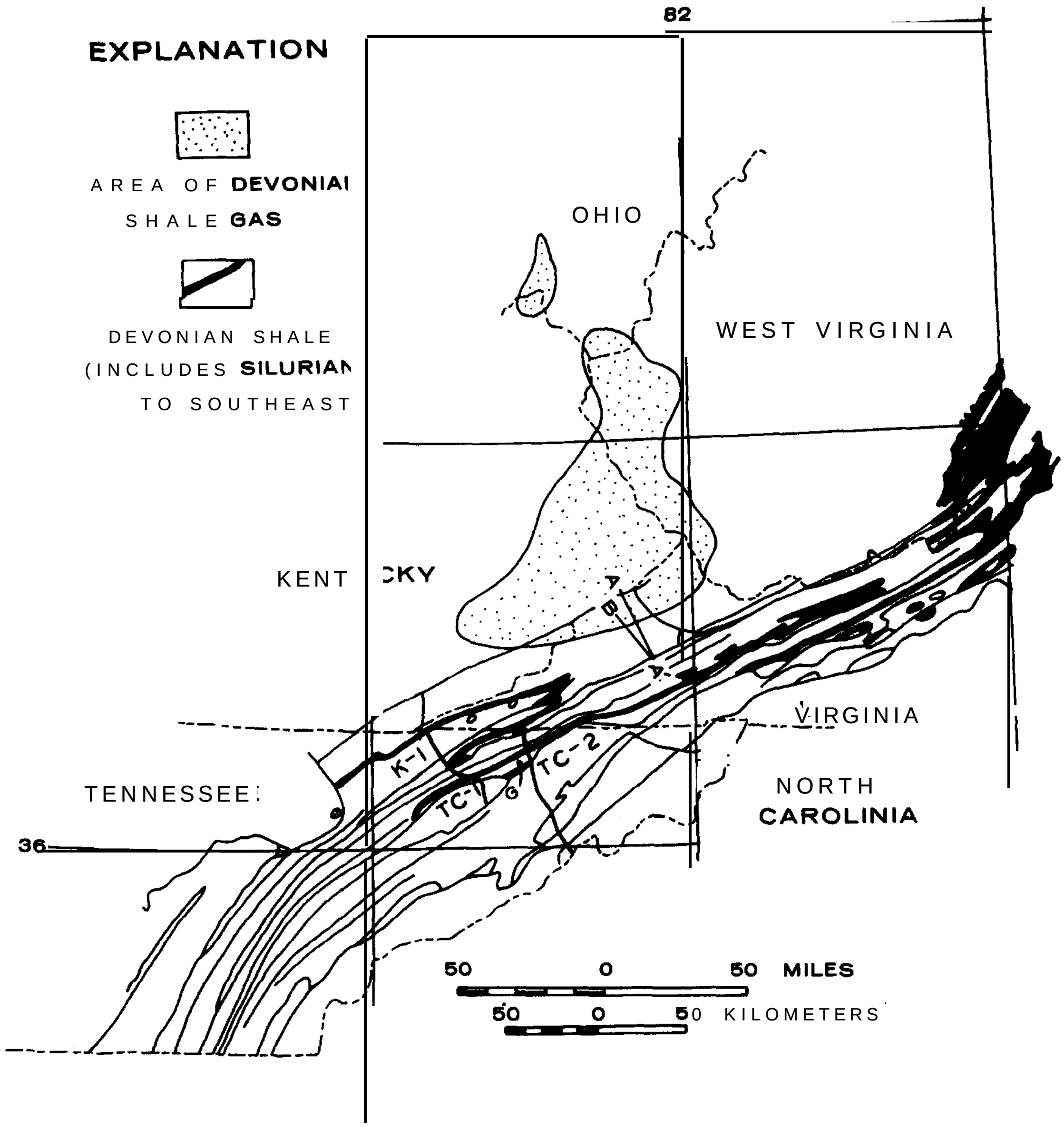


FIGURE 1. Generalized geologic map showing major faults, outcrop of Devonian black shale (including Silurian formations to southeast) and areas of production of shale gas (stippled), after King and Beikman (1974) and Wallace and deWitt (1975). Cross sections AA' and BA' are in Figure 6. For definitions of other letter symbols, see Figure 7 and text.

FIGURE 2. Tectonically fractured Chattanooga Shale at Sneedsville, Tennessee.

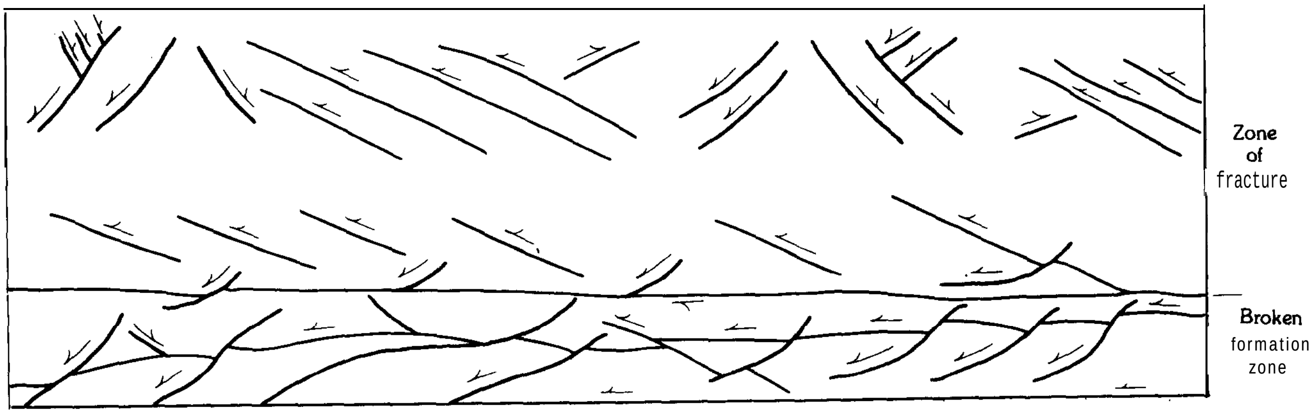


FIGURE 3. General model for mesoscopic deformation observed in southern Appalachian thrust sheets (from Harris and Milici, 1977, plate 5).

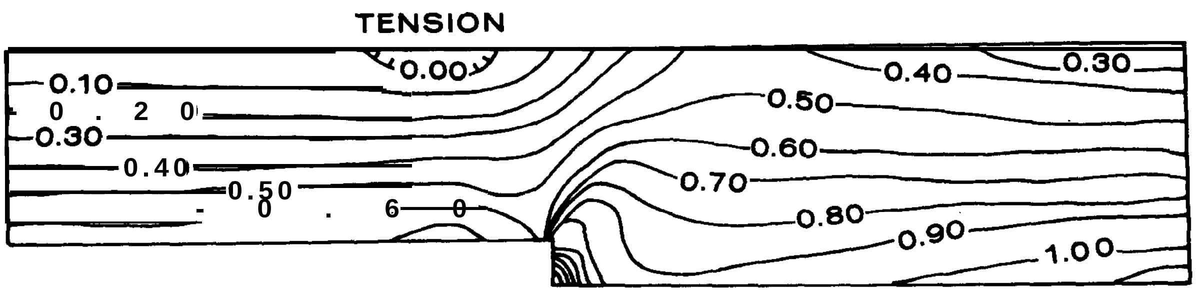


FIGURE 4. Horizontal stress contours, finite element model for single basement faulting analysis (Advani, GangaRao, Chang, Dean and Overbey, 1977, Figure 7a).

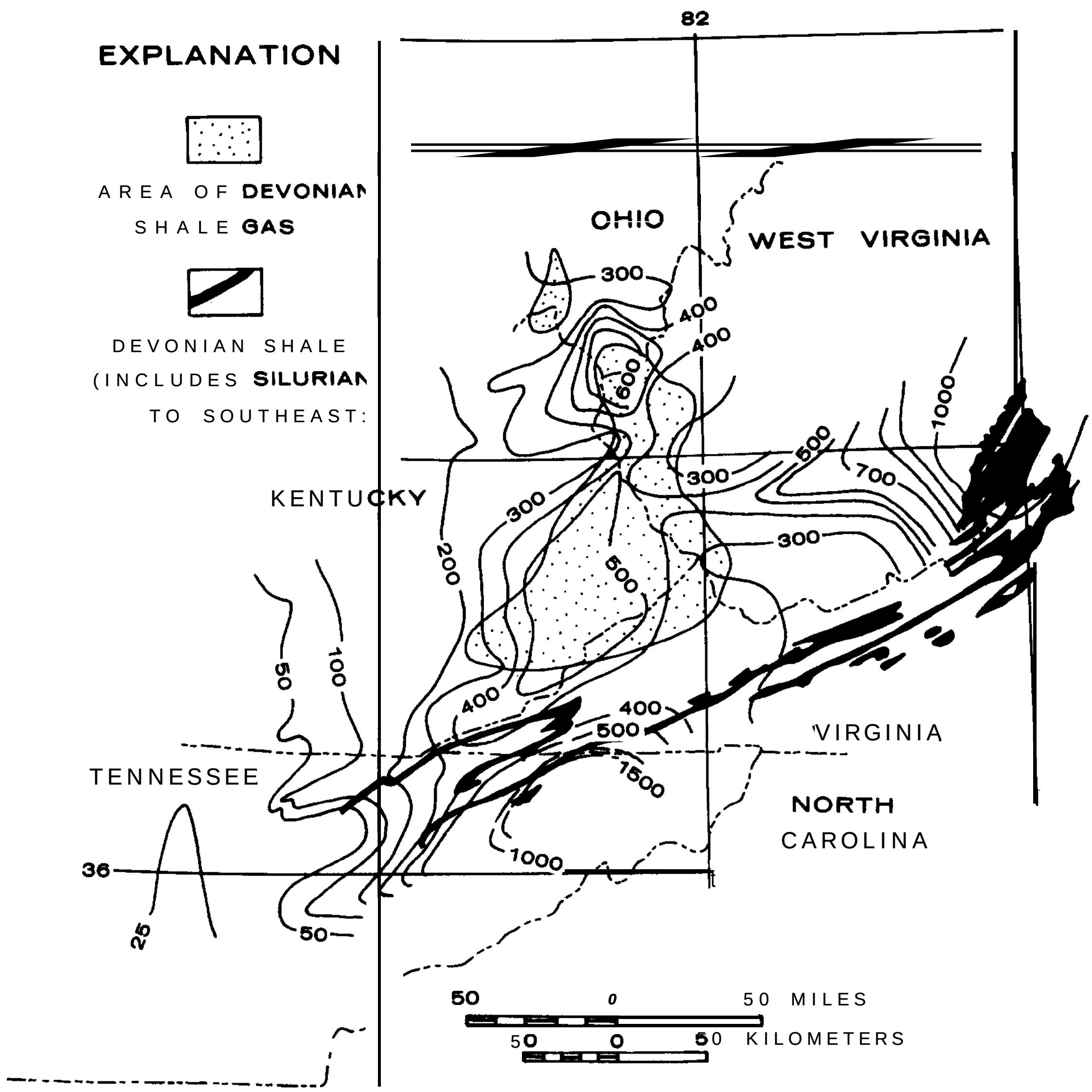
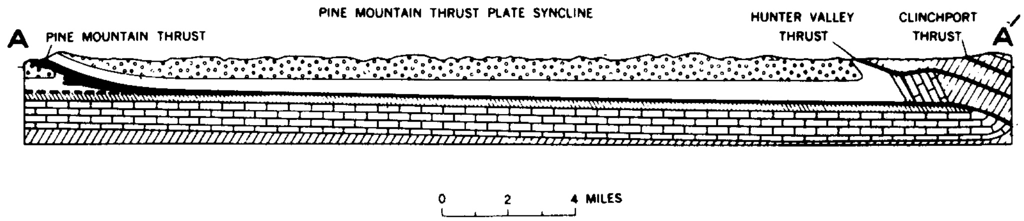


FIGURE 5. Generalized isopach map of the Chattanooga Shale in eastern Tennessee and nearby states (after Wallace and deWitt, 1975).



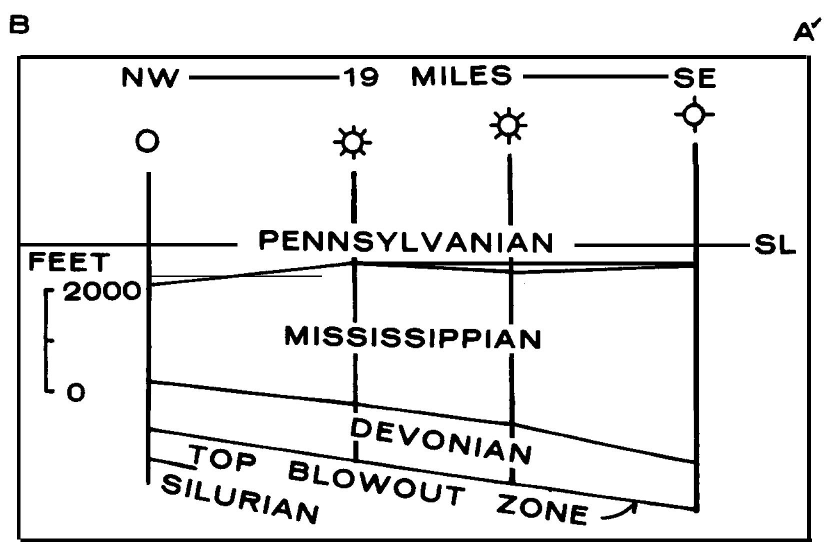


FIGURE 6. Interpretation of the blowout zone beneath the Pine Mountain block in terms of the general model for mesoscopic deformation. The blowout zone is a broken formation. A zone of fracture probably lies in the Devonian beds above. See Figure 1 for location of cross sections (cross section AA' from Harris, 1970; cross section BA' from Young, 1957).

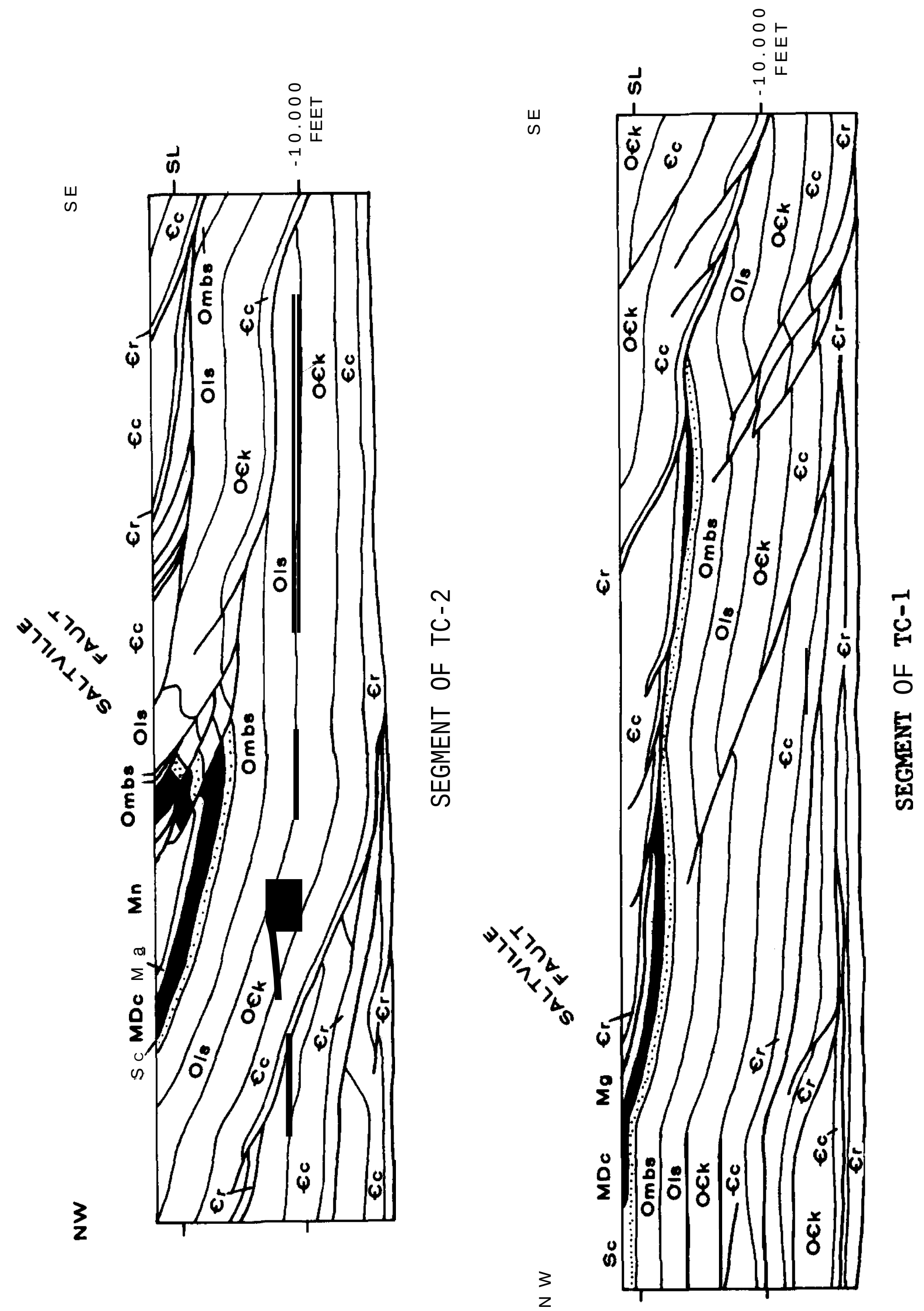


FIGURE 7. Interpretation of seismic cross section segments across the Greendale syncline (G in Figure 1) in eastern Tennessee (from Milici, Harris, and Statler, in preparation). See Figure 1 for location.

WEST-TO-EAST (BREAK-BACK) IMBRICATION OF THE ALLEGHANIAN ALLOCHTHON IN THE SOUTHERN APPALACHIAN PLATEAU AND VALLEY AND RIDGE

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ABSTRACT

The formation and distribution of southern Appalachian low angle thrusts is controlled by regionally developed lithotectonic units. A lower lithotectonic unit of terrigenous clastics is the site for lower decollement. Overlying carbonates are cross cut by tectonic ramps, and an upper terrigenous clastic unit contains upper level decollements. An analysis of fault patterns, confirmed both by the general absence of rotational effects along closely spaced faults, and by structural truncations in seismic cross sections, indicates that the sequence of imbrication of the Alleghanian allochthon was from west to east. Regional distribution of strata on the hanging walls of the major thrusts, together with data from seismic cross sections , is evidence that the basal decollement migrates upward from west to east across the Valley and Ridge.

INTRODUCTION

In the southern Appalachians closely spaced imbricate thrusts extend from near the northern end of the Saltville fault in Virginia, southwestward across Tennessee into Georgia and Alabama, where they are overridden by the westward bend of the low-angle Rome thrust block (Fig. 1). In an earlier paper I showed how patterns of overlapping thrust blocks and fault intersections suggested that an Alleghanian allochthon, sliding above a regional decollement in Cambrian strata, broke first along its toe in the Cumberland Plateau and then imbricated from west to east, from the Plateau across the Valley and Ridge to the Blue Ridge (Milici, 1975). It is the purpose of this paper to present additional lines of reasoning in support of this thesis, and to show how this sequence of fracturing can be assembled into a general model for Alleghanian deformation.

An understanding of the sequence of imbrication bears directly on interpretations concerning the mechanism of thrust implacement (gravity sliding vs. tectonic compression) and on the ultimate source of the deformational forces, i.e., continental collision vs. subduction (Milici, 1975). Furthermore, interpretations of structural patterns in the southern Appalachians may provide a basis for interpreting structural patterns in other mountain ranges.

RELATIONSHIP OF STRUCTURE TO REGIONAL FACIES

The Paleozoic stratigraphic sequence in the southern Appalachians can be divided into three gross regionally developed lithotectonic units (Fig. 2). Lithotectonic units are groupings of stratigraphic sequences that have yielded collectively and in the same way to the forces of deformation (Jacobeen and Kanes, 1974, 1975). These gross lithotectonic units control both the formation and the distribution of an integrated system of decollements and tectonic ramps throughout the southern Appalachians (Milici and Harris, 1976; Harris and Milici, 1977).

The lower lithotectonic unit is composed of terrigenous clastics of the Rome Formation and Conasauga Group (Cambrian). The middle lithotectonic unit consists mostly of carbonates that range in age from Cambrian to Lower Ordovician along the eastern side of the Valley and Ridge, and from Cambrian to Mississippian along the western Valley and Ridge and in the Plateau. The carbonate sequence is in turn overlain by an upper lithotectonic unit that is composed mostly of terrigenous clastics. These clastics range in age from Middle Ordovician on the east to Pennsylvanian on the west. Both upper and lower terrigenous lithotectonic units are sites for regional decollements, and these are connected by tectonic ramps that cut diagonally across the intervening carbonate unit.

Seismic cross sections (for example Harris, 1976; Snelson, 1976; Milici, Harris and Statler, in preparation) show that a subhorizontal master decollement lies within the lower lithotectonic unit beneath both the Valley and Ridge and the Plateau. In the Plateau, ramps and upper level decollements are developed in subhorizontal strata. In contrast, in the Valley and Ridge ramps cut across previously folded carbonates, and upper level decollements follow previously tilted beds (Fig. 3).

RELATIVE AGES OF THRUSTS

The relative ages of faults (defined as their time of last large scale movement) can be determined from an analysis of intersecting fault patterns and the geometry of contiguous fault blocks. In addition, the rotational effects (or absence thereof) of inclined tectonic ramps on their neighbors can be used to establish sequences of movement of both intersecting and non-intersecting'but closely spaced thrusts.

Intersecting fault patterns displayed at the present level of erosion are either in the upper decollement position or are along tectonic ramps. In an earlier paper I classified fault intersections into four major types, A, B, T, and X (Milici, 1975). A and T intersections develop where older thrusts are overridden and buried from the east by the next succeeding thrust block, so that the thrusts intersect at an acute angle (A-intersection) or nearly at right angles (T-intersection). The X-intersection forms where cross faults cut across the strike of the Appalachians, thereby off-setting the traces of the major Appalachian thrusts. B-intersections are hypothetical and are supposed to form where younger faults rise from beneath and off-set older faults.

A-intersections are the dominant type of fault intersection in the southern Appalachians, and these can be classified into four groupings (Fig. 4). A-intersections form where tectonic ramps intersect at an angle (Fig. 4A), where thrusts rupture along the zone of inflection at the tops of ramps as they flatten into clastics of the upper lithotectonic unit (Figs. 4B and 4C), and where upper level low angle thrust blocks ride subhorizontally one over another (Fig. 4D).

From an analysis of A-type fault intersections , where older faults are overridden and buried from the east by the next succeeding thrust block, it is evident that movement of the overridden blocks must have ceased across the Valley and Ridge in a west-to-east sequence.

If instead, an east-to-west imbrication had occurred then younger faults would either rise from beneath and off-set older faults (thereby forming B-type fault intersections, Fig. 5A), or non-intersecting faults would rise to the surface somewhere to the west of the older thrust faults. Movement along these hypothetical younger faults would result in piggy-back transportation and progressive rotation of the older thrust blocks as a succession of younger faults ramped upward on the west (Figs. 5, 6).

Rotational effects that are evidence of an east-to-west sequence of thrusting would be best displayed where major thrusts intersect along the ramp position so that piggy-backed fault blocks would be steepened by movement up the ramp (Fig. 5E). In this case, as the fault intersection is approached along the strike of the more easterly fault block, there would be a marked increase in the dip of the hanging wall formations. In examples (such as near the intersection of the Wallen Valley and Clinchport faults) oversteepening is not observed, further evidence that rotational processes are not operative and that thrust blocks are not piggy-backed by movement along faults to the west. If thrust blocks have not been rotated, then local steepening is related to other factors, such as the cross cutting of previously inclined (rather than subhorizontal) footwall beds. From the foregoing we can conclude that both evidence from fault intersections and the absence of rotational effects along closely spaced thrust faults supports the concept of a west-to-east sequence of imbrication in the thrust faulted southern Appalachian Plateau and Valley and Ridge.

EVIDENCE FROM SEISMIC CROSS SECTIONS

A general theory for Alleghanian deformation of the southern Appalachians must consider both the relative sequence of implacement of thrust blocks and the apparent upward migration of detachment within the lower lithotectonic unit. Harris (1976) in his interpretation of a seismic profile across part of the western Valley and Ridge of Tennessee showed that from west to east thrust faults form in progressively higher stratigraphic levels within the basal decollement. This upward migration of thrust faults within the Cambrian clastic sequence results in a general eastward thickening of the decollement zone'as large masses of the Rome Formation and Conasauga Group are abandoned in the subsurface (Harris and Milici, 1977).

This observation is confirmed by analysis of the regional distribution of formations along the hanging walls of the major thrusts (Fig. 7). With minor exception, the Rome Formation is prominent along hanging walls of major thrusts to the west, and the Conasauga or Knox along hanging walls of major thrusts to the east, additional evidence that the stratigraphic position of detachment rises from west to east beneath the Valley and Ridge.

A survey of available seismic data does not show structures deep in the tectonic pile rising from beneath and deforming overlying thrust sheets. Instead, along a segment of line TC-2 in eastern Tennessee the Saltville fault clearly truncates earlier formed structures beneath it (Milici, Harris and Statler, in preparation) (Fig. 6,CC').

We can conclude, therefore, that the older thrusts are deeper in the tectonic pile and rise to the surface farther to the west than the younger faults. Cross sectional shapes of the thrusts are concave upward, so that the Alleghanian allochthon appears to consist of a series of elongate nested slabs. Observable structure consists of subhorizontal upper level decollements on the west in the Plateau, a series of closely spaced tectonic ramps and inclined decollements in the western and central parts of the Valley and Ridge, and subhorizontal upper level decollements along the eastern side of the Valley and Ridge and along the toe of the Blue Ridge (Fig. 8).

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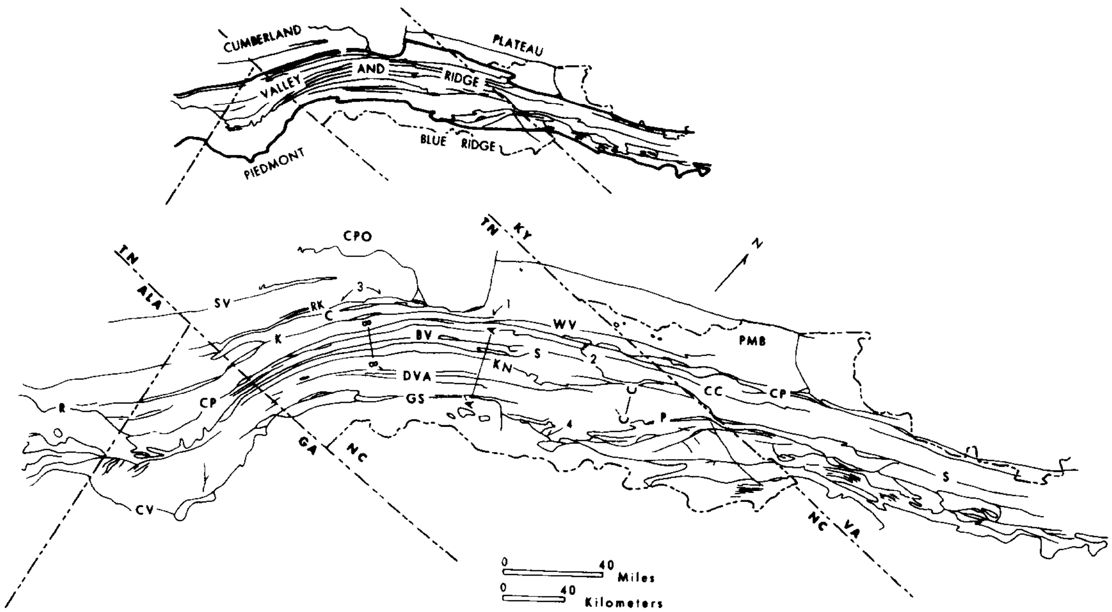
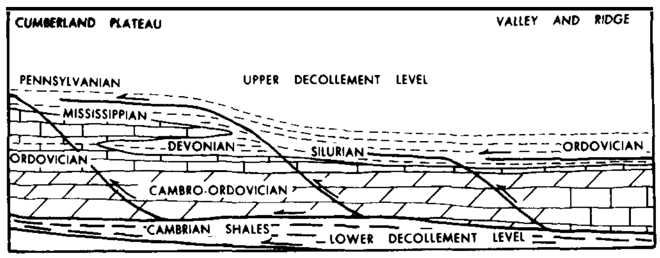


FIGURE 1. Map of major thrusts in Southern Appalachian Plateau and Valley and Ridge. Physiographic provinces are shown in inset. CPO-Cumberland Plateau overthrust; PMB-Pine Mountain block; SV-Sequatchie Valley fault; RK-Rockwood fault; C-Chattanooga fault; WV-Wallen Valley fault; K-Kingston fault; CP-Clinchport fault; CC-Copper Creek fault; R-Rome fault; BV-Beaver Valley fault; S-Saltville fault; KN-Knoxville fault; DVA-Dumplin Valley anticlinorium; P-Pulaski fault; GS-Great Smoky fault; CV-Cartersville fault. Numbers refer to examples of fault intersections described in Fig. 4.

FIGURE 2. Generalized relationship of regional facies to major decollement levels and cross cutting tectonic ramps in Southern Appalachians.



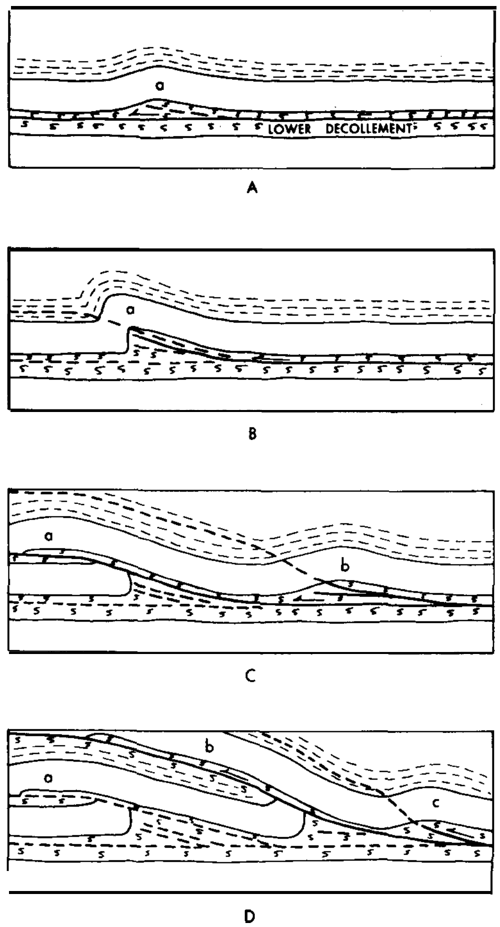


FIGURE 3. Generalized cross sections showing sequence of break-back deformation in Plateau and Valley and Ridge. A, linear folds, such as Powell Valley and Sequatchie Valley anticlines, develop at (a) above lower decollement; B, fold steepens and is faulted; C, fault cuts from lower decollement across carbonate ramp into upper decollement shales, footwall beds are flat, hanging wall beds are inclined, next fold starts to develop at (b); D, thrust at (b) cuts across inclined ramp beds and enters inclined upper decollement-level shales, next fold starts to develop at (c). Adapted from Harris (1976, Fig. 5).

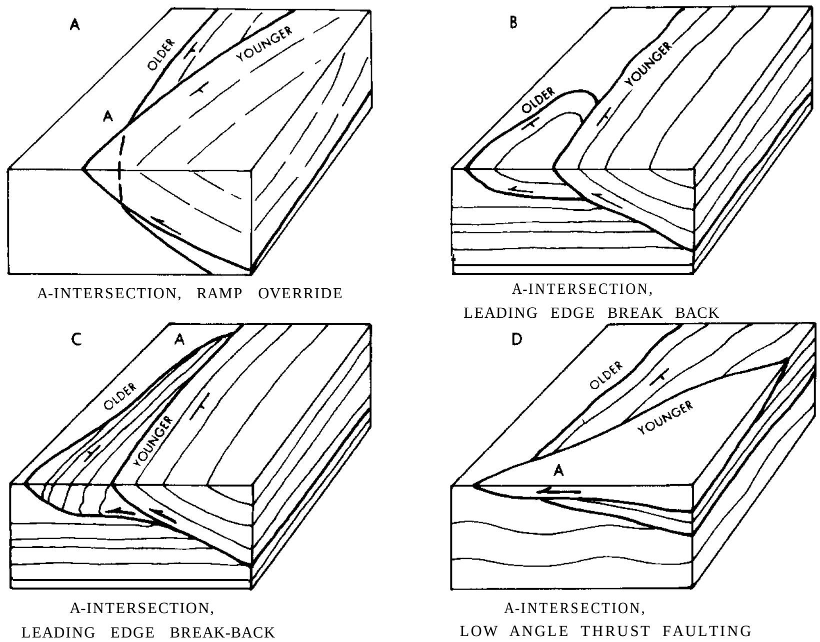


FIGURE 4. Varieties of A-type fault intersections. A, example is intersection of Clinchport and Wallen Valley faults (Fig. 1, location 1); B, example is the relationship of Hunter Valley to Clinchport fault (Fig. 1, location 2); C, example is the relationship of the Rockwood to the Chattanooga fault (Fig. 1, location 3); D, example is relationships of several thrusts near intersection of Pulaski and Great Smoky faults (Fig. 1, location 4).

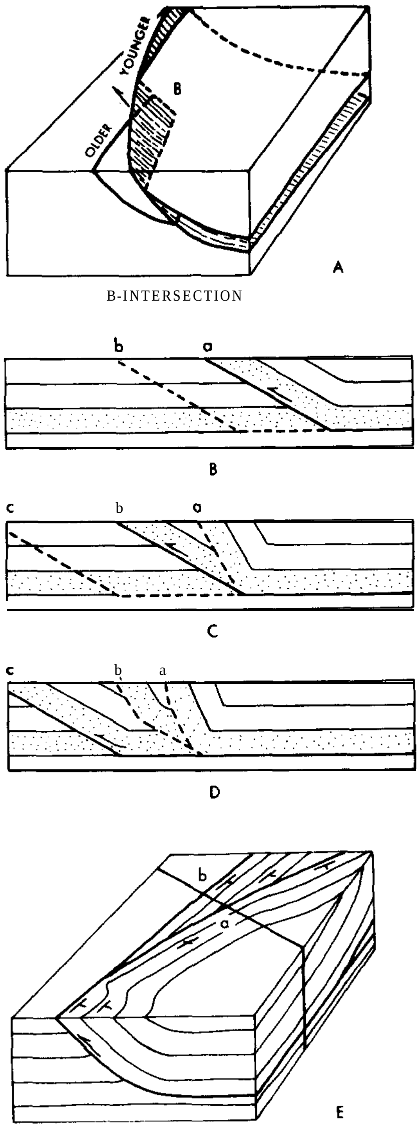


FIGURE 5. A, block diagram of hypothetical type B-fault intersection, no examples known; B, C, D, hypothetical east-to-west sequence of faulting showing cumulative effects of rotation of fault block (a) by piggy-back movement along (b) and (c); E, block diagram showing steepening of beds that must take place where faults intersect along tectonic ramps, compare with map pattern of intersection of Wallen Valley and Clinchport faults (Hardeman and others, 1966).

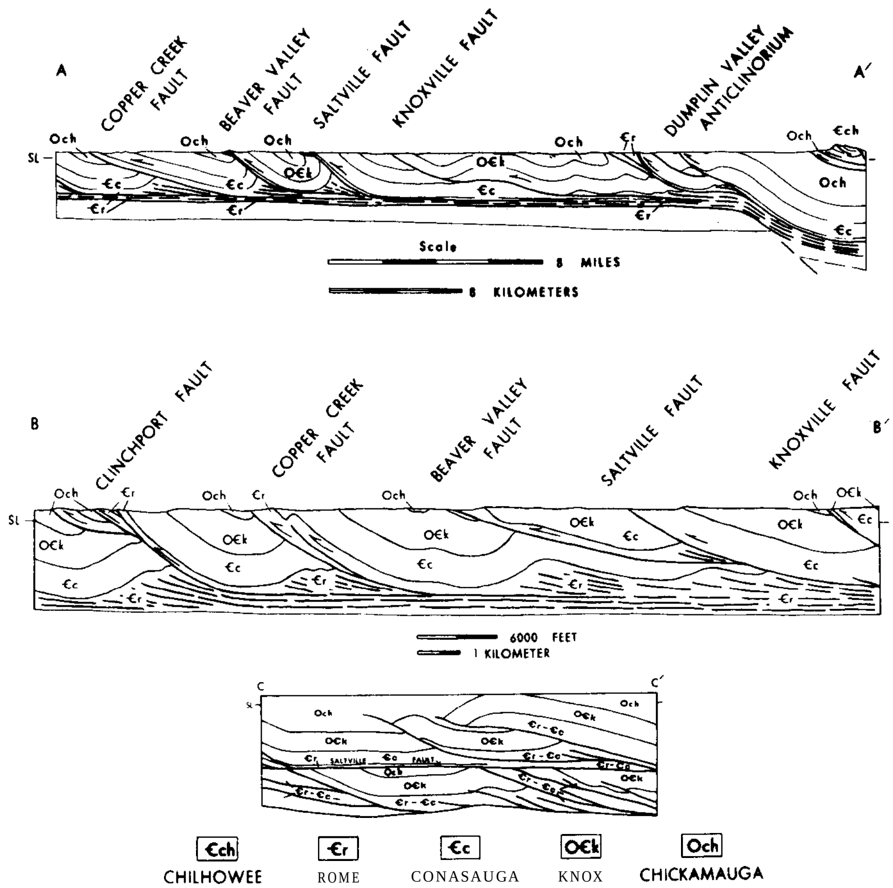


FIGURE 6. Cross section AA', shows possible rotation of Beaver Valley and Saltville fault blocks by movement of Copper Creek fault block up ramp; however, Knoxville fault block shows no effects of rotation. Cross section BB' shows absence of rotational effects in closely spaced thrusts (Modified from Rodgers, 1952). Cross section CC' is an interpretation of a segment of seismic line TC-2 (Milici, Harris and Statler, in preparation). See Figure 1 for location of sections. Scale is same for BB' and CC'.

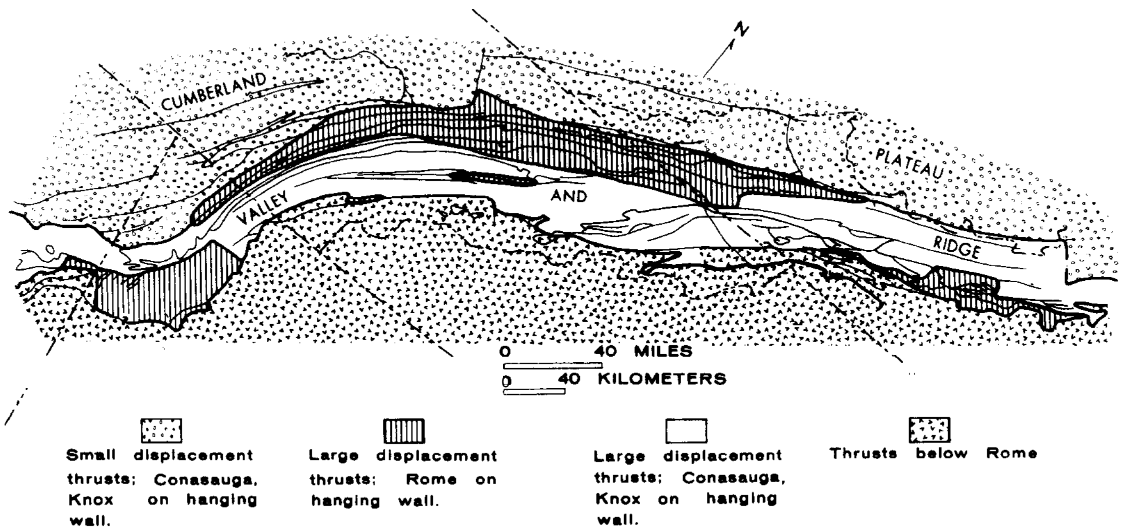


FIGURE 7. Diagram showing distribution of oldest formations on hanging wall of thrusts in Plateau, Valley and Ridge and Blue Ridge of Tennessee and parts of adjacent states.

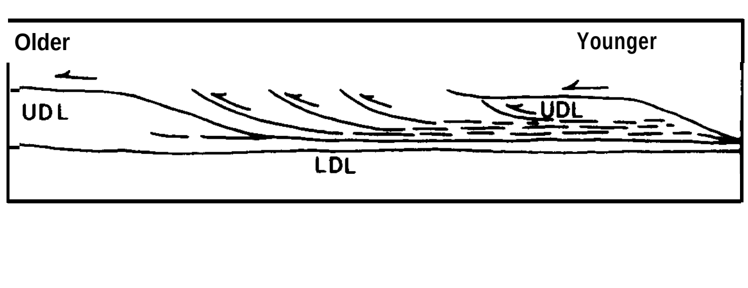


FIGURE 8. Generalized diagram showing sequence of faulting and upward migration of decollement from the Plateau to Blue Ridge in the Southern Appalachians.

LDL, lower decollement level; UDL, upper decollement level.